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Solar radiation

By C. W. ALLEN

University of London Observatory, Mill Hill Park, London, N.W.7

Symons Memorial Lecture, delivered 2 April 1958

1. Introduction

In order to appreciate fully the significance of your invitation to me this evening, I have made a point of listing the Symons Memorial lecturers and their subjects for the past twenty years. I am indeed pleased to add my lecture to the impressive list. I notice the names of some astronomers and hence it is evident that you are willing to listen to those whose main interest in the earth's atmosphere is to do away with it. Many astronomers study the earth's atmosphere with great care to the end that they shall be able to present their observational results exactly as they would have been had there been no atmosphere at all. I hold to this attitude to some extent myself. If you are willing to make due allowance for this I think you could find my subject of ' solar radiation ' well suited to the Symons Memorial series.

As all know, solar radiation is the sun's most generous gift to the earth. It provides energy to make the atmosphere work – thus it gives us our meteorology. Most of our physical knowledge of the sun is obtained from detailed studies of its radiations and therefore my subject could embrace the whole of solar physics. However, the interests of the present audience suggest that I restrict my subject matter in the following two ways; (a) by neglecting all the detailed surface phenomena of the sun and considering only the totality of radiation covering the whole of the sun's visible disk and (b) by neglecting all the detail of the Fraunhofer spectrum lines and considering only a spectrum that is very much smoothed in wavelength. Another restriction you might expect of me is that I should consider only total radiation integrated over the whole spectrum, but the influences of the solar radiation on the earth's atmosphere are so dependent on general wave length (although not on the detailed spectrum lines) that the distribution of energy in wavelength could not be omitted.

These introductory remarks guide my discussion into three main questions. How much radiation comes from the sun ? How does it vary ? How much gets lost in the atmosphere ? You will have noticed that another important question is deliberately omitted. I do not attempt to explain how the solar radiation affects the atmosphere – a question which might well be left to meteorologists.

In order to express results quantitatively we might denote by f_λ the flux of radiant energy from the sun per wavelength unit reaching unit area outside the earth atmosphere. Using the unit, erg cm^{-2} $Å^{-1}$ sec^{-1}, the maximum f_λ is about 216, near $\lambda = 4700$ Å. If we should wish to represent the flux of the continuum *between* the spectrum lines this will be denoted f_λ'.

It will be necessary to consider f_λ over a very wide range of wavelength (10^{-7} to 10^3 cm) and to do this in a single diagram (as in Fig. 3) logarithmic wavelengths must be adopted. However if f_λ be plotted against $\log \lambda$ the area under the curve so produced does not represent total energy over the wavelength range. As total energy is so often our main concern, and as $\log \lambda$ is a very useful abscissa (even for ranges in λ as small as 1 : 100) I have introduced the symbol $f_\%$ to represent the flux in ergs per cm^2 per sec and per 1 per cent change in wavelength or frequency. $f_\%$ is a more useful quantity than f_λ for expressing the spectral flux when a wide wavelength range has to be considered. We have

$$f_\% = \lambda f_\lambda / 100$$

and the total energy flux over a wavelength range can be written

$$f = \int f_\lambda \, d\lambda = 230 \int f_\% \, d(\log \lambda)$$

where the numerical factor 230 is 100/modulus, and logarithms are to the base 10.

2. The permanent spectrum

It will be convenient to consider the whole spectrum in three parts : (a) the visible spectrum which is to include also the near infra-red and ultra-violet, (b) the radio spectrum and (c) the x-ray and far ultra-violet spectrum.

The visible spectrum together with the near infra-red and ultra-violet (0·2 μ to 10 μ) contains virtually all the solar energy. Much of it reaches the earth's surface and there it can be measured by various types of thermal radiometer. The evaluation of f_λ in this region is dependent on the absolute calibration of radiation instruments and on appropriate allowance for absorption in the earth's atmosphere. Rocket observations are required

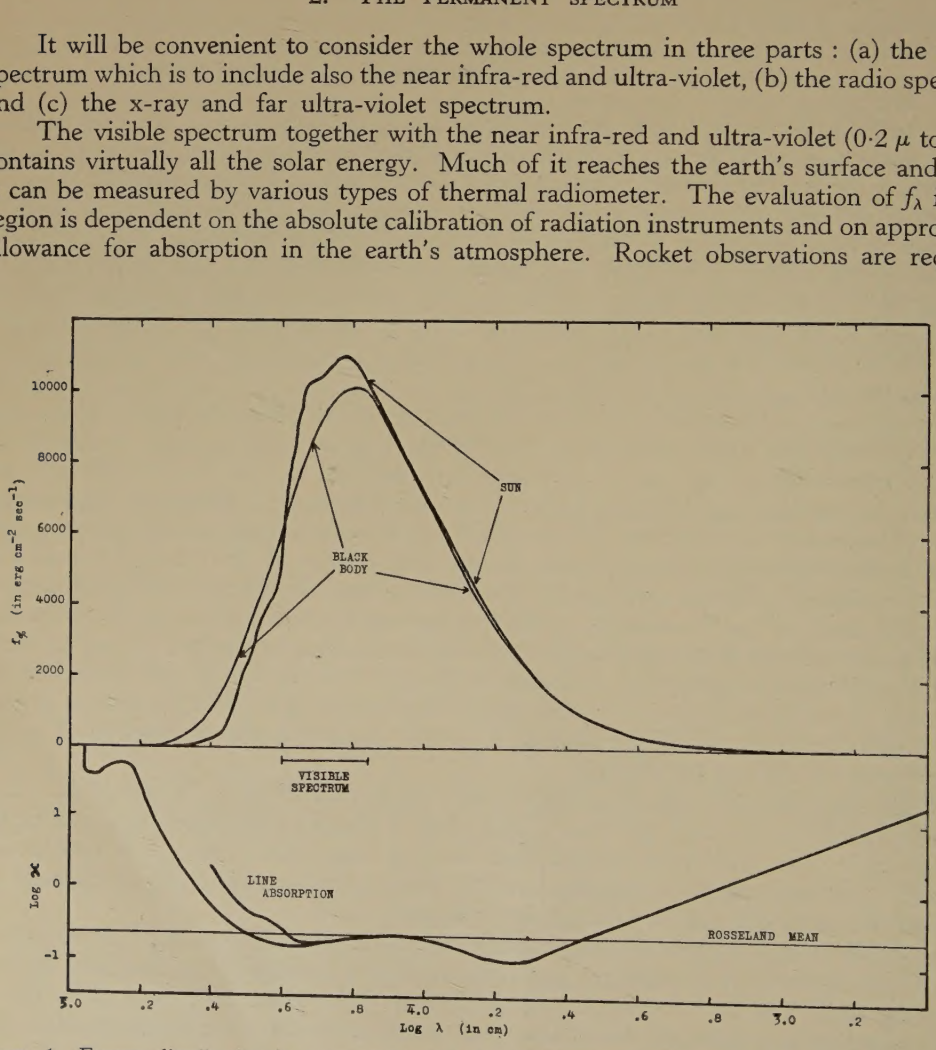

Figure 1. Energy distribution in sun and black body. Absorption coefficient κ of the solar atmosphere.

in the range 0·2 μ to 0·3 μ. In Table 1 I have reconsidered earlier data (Allen 1955) with the help of discussions by Deirmendjian and Sekera (1956), Johnson, Purcell, Tousey and Wilson (1954), and Johnson (1954). Some account has also been taken of measurements of the intensity of the centre of the sun's disk in the continuum between Fraunhofer lines by Makarova (1957), Peytureau (1952) and Labs (1957). Such measurements can be converted to f_λ by a knowledge of solar limb darkening and the total absorption of Fraunhofer lines, and available estimates of these have been used (extrapolated to some extent) for compiling Table 1. In this table the wavelengths are chosen to show the variation in as much detail as seems necessary. I have used λ, f_λ and log $f_\%$ in the table, but in graphical work I use log λ and $f_\%$. The derived energy distribution is shown in Fig. 1 and compared with a black-body distribution for the sun's effective temperature T_e which is 5800°K. The area under the $f_\%$ curve is proportional to total energy, and by definition of T_e the areas under the solar and black-body curves are the same. Integration of the f_λ data of Table 1 gives $1{\cdot}380 \times 10^6$ erg cm^{-2} sec^{-1} $\equiv$ 1·978 cal cm^{-2} min^{-1}.

The departure of the solar distribution from a black-body curve, though small, is significant and interesting. The lower part of Fig. 1 shows the value of κ the opacity of the sun's atmosphere per (g/cm^2). Where κ is less than the mean value (represented in the diagram as the Rosseland mean) we have a 'window' which enables us to look more deeply than average into the sun's atmosphere. As the sun's temperature rises steeply with depth, the emission in such a window is represented by a black body with temperature higher than T_e. Two windows are shown : one in the visible and one in the infra-red beyond $\lambda = 1\ \mu$. In both cases the emission is greater than the black body, but the effect is much less sensitive in the infra-red than in the visible. Since our f_λ and $f_\%$ refer to the spectral flux smoothed through spectral lines the lines themselves must be considered to add to the opacity of the sun's atmosphere. The extra curve in Fig. 1 is intended to represent this extra line absorption. Because of it we see that in the near ultra-violet the absorption is greater than the mean. In this region we are therefore looking at a cooler solar level (nearer the surface) where the radiation intensity is less than black body at T_e.

It is evident that the form of f_λ is dependent on the shape of the κ curve which, as we see in Fig. 1, rises steeply towards either longer or shorter wavelengths. How this will affect f_λ will depend on the distribution of temperature with height in the sun's atmosphere. This is illustrated in Fig. 2. The critical part of the curve near the photosphere is well determined and the temperature in the corona is known to be close to 10^6 °K as shown. However the shape of the temperature curve between the photosphere and corona has been subject to very considerable controversy and there are strong reasons for thinking that in this region there are pockets of high and low temperature material

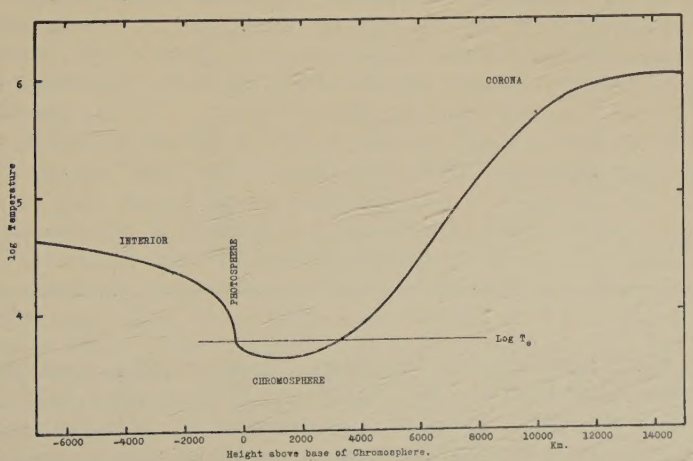

Figure 2. Temperature distribution in the sun's atmosphere.

that create a complex chromospheric structure. The point for us to notice is that there is a level of very high temperature above the chromosphere, and hence if the absorption coefficient of the atmosphere becomes very great the characteristic black-body temperature may be much greater than T_e.

TABLE 1. SOLAR RADIATION FLUX OUTSIDE THE EARTH'S ATMOSPHERE

Visible spectrum, etc.

λ μ	f_λ (erg cm^{-2} Å^{-1} sec^{-1})	log $f_\%$ (in erg cm^{-2} sec^{-1}
0·20	0·5	1·0
0·22	3·0	1·82
0·24	6	2·16
0·26	12	2·49
0·28	24	2·83
0·30	55	3·22
0·32	75	3·38
0·34	97	3·52
0·35	105	3·57
0·36	109	3·59
0·37	112	3·62
0·38	115	3·64
0·39	123	3·68
0·40	153	3·79
0·41	176	3·86
0·42	186	3·89
0·44	203	3·950
0·46	215	3·995
0·48	214	4·012
0·50	206	4·013
0·55	195	4·030
0·60	183	4·041
0·65	164	4·027
0·70	146	4·009
0·75	128	3·982
0·80	113	3·957
0·9	89	3·90
1·0	72	3·86
1·1	59·5	3·82
1·2	49	3·77
1·4	32·5	3·66
1·6	22·3	3·55
1·8	15·2	3·44
2·0	10·8	3·33
2·5	4·97	3·09
3	2·63	2·90
4	0·93	2·57
5	0·41	2·31
6	0·21	2·10
7	0·12	1·92
10	0·023	1·36

Radio spectrum

λ (cm)	f_λ (erg cm^{-2} sec^{-1})	log $f_\%$ (in erg cm^{-2} sec^{-1})
0·6	2·8 × 10^{-5}	− 6·77
1·5	1·1 × 10^{-6}	− 7·78
3	9·7 × 10^{-8}	− 8·54
6	1·1 × 10^{-8}	− 9·18
15	7·7 × 10^{-10}	− 9·94
30	1·1 × 10^{-10}	− 10·48
60	1·8 × 10^{-11}	− 10·97
150	9·1 × 10^{-13}	− 11·87
300	6·0 × 10^{-14}	− 12·74
600	4·2 × 10^{-15}	− 13·60

X-ray spectrum, etc.

λ Å	f_λ (10^{-5} erg cm^{-2} Å^{-1} sec^{-1})	log $f_\%$ (in erg cm^{-2} sec^{-1})	log $f'_\%$ (in erg cm^{-2} sec^{-1})
10	4	− 5·4	− 5·4
20	23	− 4·3	− 4·5
50	110	− 3·3	− 3·8
70	120	− 3·1	− 3·7
100	70	− 3·2	− 3·8
200	11	− 3·7	− 3·8
300	10	− 3·5	− 4·4
500	8	− 3·4	− 4·4
700	10	− 3·2	− 3·8
800	20	− 2·8	− 3·0
900	60	− 2·3	− 2·3
1000	40	− 2·4	− 2·8
1100	90	− 2·0	− 3·0
1200	280	− 1·5	− 2·8
1400	540	− 1·1	− 1·6
1600	1600	− 0·6	− 0·7
1800	10000	+ 0·2	
2000	50000	+ 1·0	

In Fig. 3 a much more compact logarithmic scale is used to show the distribution of $f_\%$ over the whole range from $\lambda = 10^{-7}$ to 10^3 cm. By comparison with the black-body spectrum at T_e there are seen to be superimposed appendages at both the radio and x-ray ends. These are both due to the high temperature corona which is completely transparent to the visible spectrum but is somewhat opaque to certain radio and x-ray wavelengths.

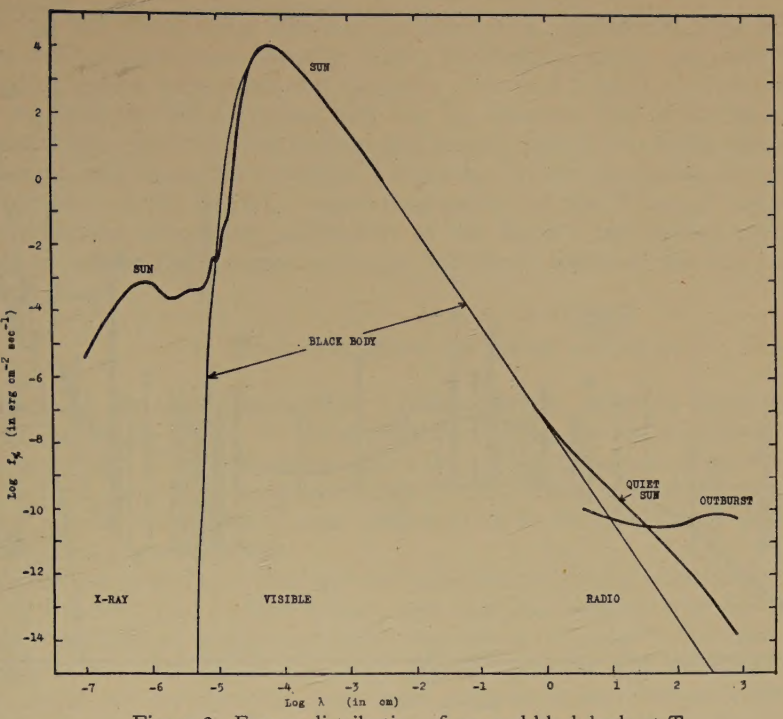

Figure 3. Energy distribution of sun and black body at T_e.

In the radio wavelengths the most intense radiations, such as those from outbursts, are neither thermal nor permanent. The permanent thermal (quiet sun) radiations can be measured by radio techniques with considerable absolute accuracy but these too vary with solar activity. The values quoted in Table 1 represent the permanent (but not precisely constant) component after the radiation from active areas has been subtracted (Allen 1957a, 1957b). The energies involved are vanishingly small. We see, for example, that the energy received from wavelengths greater than 10 cm is of the order 10^{-8} erg cm^{-2} sec^{-1}, as compared with $1{\cdot}39 \times 10^6$ erg cm^{-2} sec^{-1} for the total radiation.

The x-ray and far ultra-violet regions cannot penetrate the earth's atmosphere and therefore these radiations cannot be measured at the earth's surface. Their importance is due to their ability to ionize the upper atmosphere and produce the ionospheric layers. There are three ways by which we estimate the flux; (i) by measurements from rockets, (ii) from solar coronal and chromospheric theories and (iii) from interpretation of ionospheric observations. None of these is particularly reliable but fortunately they show some measure of agreement. In order to evaluate the radiation we start with the coronal calculations of Elwert (1954). Values for both emission lines and the continuum are available and are shown on the left of Fig. 4. In this diagram the individual line fluxes are represented in erg cm^{-2} sec^{-1}, and therefore in order to compare them quantitatively with the continuum they should be regarded as having a width of 1 per cent in λ, i.e., (1/230) $\times$ logarithmic unit. It is found that the emission lines provide a more significant contribution than the continuum. The spectrum lines further to the right in the diagram have been either observed, (Lα) (Byram, Chubb, Friedman and Kupperian 1956a), or calculated (Woolley and Allen 1948, 1950) from models of the corona (Mg x) or the chromosphere (other lines). The calculations have been reduced by a factor 6 to allow for over-estimation mentioned by Elwert (1954). The totality of emission lines when represented as a smooth curve show approximate agreement with x-ray flux measurements made from rockets (Byram *et al.* 1956a, 1956b, Friedman 1957) but no measurements are available in the interesting region from 100 Å to 1000 Å. The curve of $f_{\odot}$ in Fig. 4 and represented in Table 1 is a compromise between calculations and measurements.

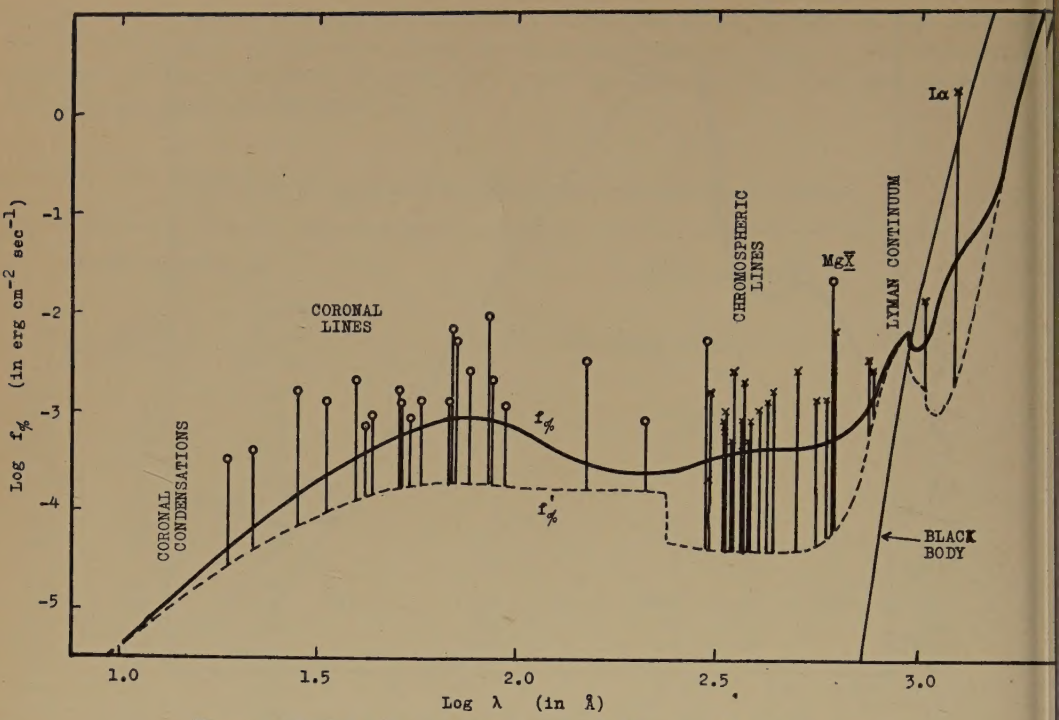

Figure 4. Emission lines and continua in the X-ray and far u-v regions.

The f_{λ}' curve for the continuum is mainly from Elwert's calculations and has not been observed. The continuum peak in the neighbourhood of the Lyman limit (912 Å) is suggested by chromospheric theory (de Jager 1955) but also has not been observed. The shortest wavelength to which the continuum has been observed is 1550 Å (Johnson, Mallitson, Purcell and Tousey 1958).

A point of considerable interest, noted by Elwert, is that the smoothed distribution representing the totality of coronal emission lines has a maximum at a wavelength close to the coronal temperature black-body maximum. In coronal condensations, where temperatures are thought to reach 6×10^6 °K, emission lines of shorter wavelength become prominent (Elwert 1956). Thus a change of emission wavelength with change of solar activity is to be expected. Fig. 4 is intended to represent quiet conditions, i.e., the sun without coronal condensations.

We might now compare energies in various parts of our spectrum with those required to form the ionospheric layers. The energies in various parts of Fig. 4 might be summarized as follows :

Spectroscopic range	x-ray	200-800 Å	Lyman Cont.	Lα
Energy (in erg cm^{-2} sec^{-1})	0·10	0·05	0·16	1·6

Comparable data for the ionosphere (Allen 1955 p. 124, 1956) are :

Ionospheric region	D	E	F
Ionization (ion pairs cm^{-2} sec^{-1})	10^5	6×10^8	25×10^8
Energy (in erg cm^{-2} sec^{-1})	10^{-5}	0·013	0·06

For these data each ionizing photon is considered to have 1 Rydberg of energy and the D region recombination coefficient is taken as 10^{-6} cm^3 sec^{-1} (Waynick 1955). There are two factors which may increase the calculated energy required to form the ionospheric

regions : (i) the ionospheric effective recombination coefficients may be greater than assumed and (ii) x-radiation has more than 1 Rydberg of energy per photon (which however might produce more than one ion pair). We find a great discrepancy between the energy required for the D region and the Lα emission that is thought to produce it (Friedman 1957). For the other regions and ranges there is reasonable agreement between spectroscopic and ionospheric energies. None are known accurately enough to help with the recognition of the spectral ranges responsible for the E and F regions. However, arguing from the absorption coefficients of the Earth's atmosphere (Allen 1956) it is plausible to ascribe the E region to x-rays and the F region to the line radiations in the 200-800 Å range.

3. The variations of solar radiation

There are certainly considerable variations in the radiation from small regions of the sun's surface as shown by the existence of sunspots, and analogous variations can be detected within certain narrow spectrum lines. But these phenomena do not necessarily affect the total radiation to any appreciable extent. Our first problem then – the problem of prime interest for meteorological purposes – is to decide whether the total radiation varies effectively or measurably.

We can see from Fig. 1 that for total radiation we need consider only the wavelength range 0·2 μ to 10 μ, and we have already seen that this radiation is almost entirely thermal in origin. Our interest relates to possible variations of the order 0·1 per cent. The observed dispersion of the variation is not much greater than this and variations that are much smaller could not be detected nor would they have any meteorological significance. Possible variations in the solar radius need not concern us since they are not much greater than 0·01 per cent (Gething 1955, Giannuzzi 1957) and have negligible effect on radiation output. A variation of 0·1 per cent in the solar constant represents a change of only 0·025 per cent or 1·5 °K in T_e.

The existence of the sunspot cycle and its superficial similarity to certain variable stars might be thought to indicate a considerable variation in solar radiation. The importance of the question led the Smithsonian Institution to embark on an extensive programme of daily measurements of the ' solar constant,' which is the term used to represent total radiation from the sun outside the earth's atmosphere expressed in calories per cm^2 per min at mean solar distance. Early observations in and around 1922 (Abbot, Aldrich and Hoover 1942) showed some variations as large as 1·5 per cent but such variations were not repeated in later years and were apparently not real. In the following years a careful search was made for any systematic relation between the solar constant and the sunspot cycle but up to 1940 no such relation could be detected. There was indeed very little evidence that any of the apparent variations were real.

The two analyses most likely to give indications of real variability are (a) a study of the covariance of measures of the solar constant made at independent stations and (b) a further study of solar constant variations with the sunspot cycle. In either case extensive and independent estimates of solar constant variations are the main requirement.

As there has been a discontinuity in the Smithsonian observing programme at the end of 1955 it would appear that it will be very many years before there are any more data available that can help with this problem and therefore the appropriate time to make a decision on this matter is *now*.

Three different estimates of the standard deviation of the solar constant by covariance methods gave the results in Table 2. The only correlation of entirely independent data (S, C) gave no evidence of covariance. The data for the two (T, M) correlations were reduced by the same institution, and there is a danger of a spurious covariance resulting from : (a) adjustment or selection of one value to improve agreement with the other, (b) meteorological correlation between the stations and (c) similarity in the residual annual term at the two stations. With regard to (c) it has been noticed that all four stations

TABLE 2. STANDARD DEVIATION OF THE SOLAR CONSTANT BY COVARIANCE AT TWO STATIONS

Period	Stations	Standard deviation	Reference
1926-1939	(T, M)	0·13 (per cent)	Sterne (1942)
1927-1939	(S, C)	0·0 (per cent) estimated; 0·1 (per cent) possible	Rimmer and Allen (1950)
1926-1955	(T, M)	0·17 (per cent)	Sterne and Dieter (1958)

T = Table Mountain (Smithsonian); M = Montezuma (Smithsonian); S = Smithsonian preferred constant; C = Canberra (pyrheliometer residual).

occupied by the Smithsonian Institution tended to minimum values in May and maxima in November. The covariance between T and M from this cause alone amounts to $(0{\cdot}07 \text{ per cent})^2$ which is an appreciable fraction of the detected covariance $(0{\cdot}13 \text{ per cent})^2$.

We might conclude that up to 1940 there was no strong evidence of detectable solar variability.

Since 1940 the Smithsonian Institution has found an apparent increase of radiation with increase of solar activity (Aldrich and Hoover 1954, p. 165). This is in such contrast with earlier findings that a search for another explanation appears to be justified. To this end I have plotted the Smithsonian solar constant (S.C) values against Zurich sunspot number R in Fig. 5. The data have been broken up into a number of discrete periods and the diagram appears to show that there have been some rather sudden changes in the Smithsonian scale, especially prior to 1927. A general change of solar constant with R can readily be seen in the whole 1942-1952 data, but if this is broken into three periods as shown there is no such correlation within each period. It would appear that the whole variation might be due to scale changes in the solar constant procedure. To check this I have sought for similar changes in the pyrheliometry grouped under selected ranges of water vapour content, pyranometry, and air mass (Aldrich and Hoover 1954, pp. 169-174). The trend of the pyrheliometry is compared with sunspots and solar constant in Fig. 6. It is seen that the maximum around 1948 is not reproduced in the pyrheliometry and this suggests that some of the solar constant changes are associated with the detailed reduction procedure.

My conclusion is that the real dispersion in total solar radiation is not greater than 0·1 per cent and that there is no justification for thinking that solar variability has been detected experimentally.

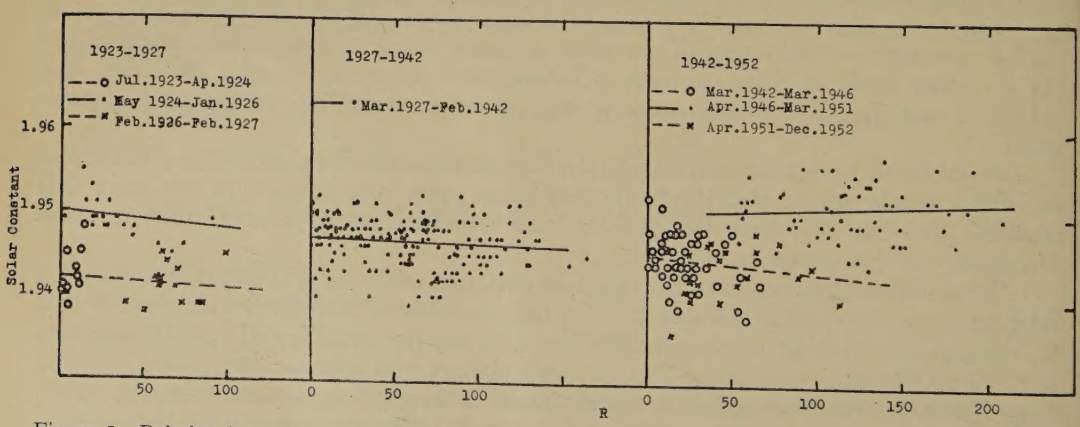

Figure 5. Relation between measured solar constant and sunspot number. The data are broken up into discrete periods.

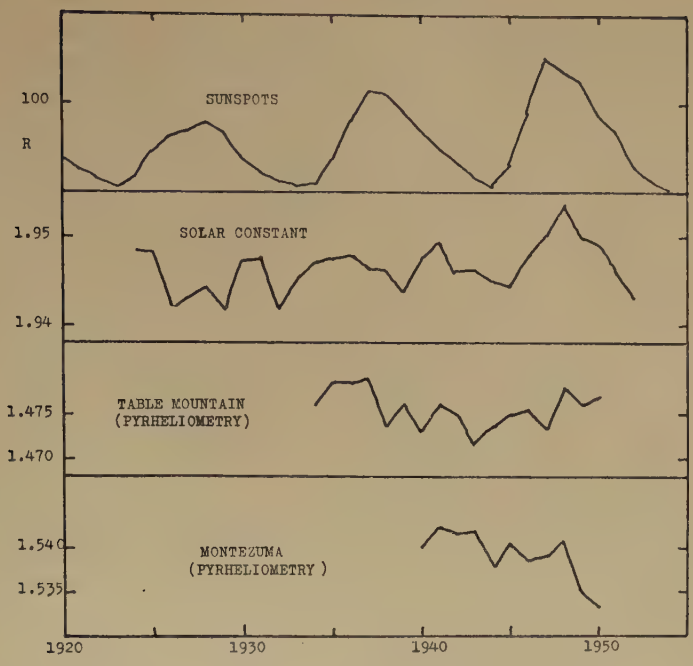

Figure 6. Comparison of sunspot numbers, solar constant, and controlled pyrheliometry.

This shows solar radiation to be quite a good constant of nature and suitable for the astrophysical radiation standard. The loss of radiation as a direct result of darkening by sunspots would amount to about 0·0009 cal cm^{-2} min^{-1} for $R = 100$. This might be compared with the drop of 0·0007 cal cm^{-2} min^{-1} detected at sunspot maximum in the 1927-42 section of Fig. 5. However the solar constant accuracy is not great enough to conclude that this sunspot loss effect has been detected experimentally.

Another extensive search for solar variability was made by Pettit (1932, 1933-1938) by making a direct comparison of radiation flux in the ultra-violet (0·32 μ) and the green (0·50 μ) regions. It is quite evident that the measurements were affected by both atmospheric variability and a secular change. Fig. 7 shows the residual variations after an

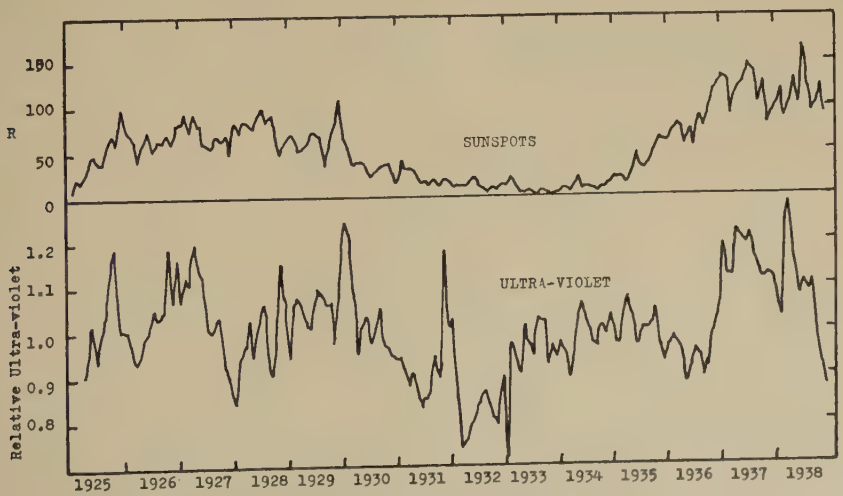

Figure 7. Comparison of sunspot numbers with ultra-violet intensity after correcting the latter for seasonal and secular variation.

annual term and a linear secular variation have been extracted. An ultra-violet intensity variation of perhaps 10 per cent in phase with the sunspot cycle would appear to be possible but the linear secular correction I have adopted gives a rather optimistic idea of the true correlation. We have seen that the radiation concerned is from thermal causes. A 10 per cent ultra-violet variation would represent a 200°K temperature change and a consequent 15 per cent solar-constant change which is certainly not observed. This leads to the conclusion, already generally agreed, that the measured variations are associated with atmospheric transmission rather than solar variation.

While the variation of total radiation is very small and uncertain the variations in radio emissions from the sun are large and show well-defined solar activity effects. The variable part can be sub-divided into slowly varying thermal emissions and three types of erratic bursts. All of these change significantly with wavelength. All come from localized regions of activity and are additional to the general thermal radio radiation of Table 1. All of them change with the sunspot cycle but only the thermal emissions from active areas (in the decimetre wave band) show the rather smooth daily and monthly variations characteristic of sunspot numbers (Allen 1957b). The bursts are more erratic. These variable radiations penetrate the atmosphere to reach the earth's surface and could possibly produce some geophysical effects. However the extreme smallness of the available energy makes this unlikely and they are probably swamped by man-made radio noise.

It was shown earlier that both the radio and the x-ray emissions were due to the hot corona. Since these may be considered as being due to the same cause one might expect that the variations would show striking similarities. This is indeed the case. The variations in the ultra-violet or x-ray emissions giving rise to the E and F layers may be estimated from ionospheric critical frequency measurements. The detailed comparison of the data (yearly, monthly and daily) of the ionospheric regions, the decimetre wave radio emission, and various solar activity indices agree with such precision (Allen 1946, 1948, 1957b; Denisse and Kundu 1957) that the association of causes is no longer in doubt. Small differences in the form and timing of these variations are now being studied for solar information on various phenomena associated with a centre of solar activity. For geophysical consideration the results can be expressed by empirical relations between the radiation and the activity indices (Minnis and Bazzard 1958).

4. Loss of solar radiation in the earth's atmosphere and the solar constant

The quantitative problem of radiation loss in the earth's atmosphere is one that is of vital interest to both astronomers and meteorologists. The main atmospheric absorbing agents that operate in the parts of the solar spectrum containing appreciable energy are listed below. For each agent one requires an absorption coefficient as a function of wavelength λ. If the absorption is continuous the radiation loss may be determined simply

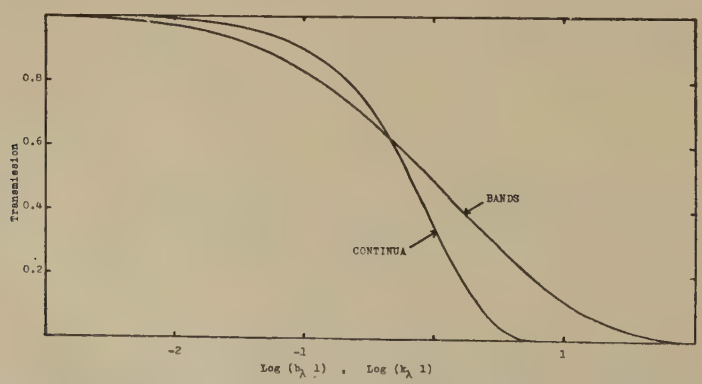

Figure 8. Comparison of transmission through bands with transmission associated with the exponential absorption of continua.

from the Lambert exponential law. If however the absorption is made up of complex lines or bands other less precise methods must be devised and an effective absorption coefficient must be defined. Let T be the transmission in a spectral region λ which is taken wide enough to include several spectrum lines. T will depend on an effective absorption coefficient b_λ and the amount of the agent which may be denoted l. Fortunately the function $T(b_\lambda l)$ is not too seriously dependent on the distribution of absorbing lines (Cowling 1950, Goody 1952, Howard, Burch and Williams 1956), and useful accuracy may be obtained from a mean curve. The curve adopted here (Allen 1955, p. 118) is compared with a curve representing exponential absorption in Fig. 8. If the value of b_λ is based on measurements near $T(b_\lambda l) \simeq 0{\cdot}5$ the amount of light absorbed will not depend very greatly on the shape of the $T(b_\lambda l)$ curve.

The main absorbing agents (Allen 1955, pp. 116, 118, 120) are as follows :

(a) Molecular (Rayleigh) scattering. This is known with a precision of about $\pm$ 2 per cent. It has a continuous spectrum following a λ^{-4} law.

(b) Ozone. Absorbing bands occurring in the ultra-violet and yellow-red regions have virtually a continuous spectrum. Absorption coefficients are known and the ozone world distribution is fairly regular.

(c) Aerosols (dust, droplets, smoke, etc.). The amount of this absorption and its dependence on wavelength may vary and hence they must be determined experimentally for each occasion. The absorption is continuous and usually decreases slowly with wavelength.

(d) Bands of water vapour and other molecules. The estimation of transmission through these complex spectrum line structures may be made by the approximate method mentioned above. Values of b_λ are available but not very precise.

The flux of solar radiation transmitted through the earth's atmosphere computed from data of the type enumerated and from Table 1 is represented in Fig. 9. Using unit air mass of aerosol-free air containing 10 mm of precipitable water I obtain a calculated total solar transmission $T_c = 0{\cdot}758$. If pyrheliometric observations representing these conditions were available the observations could be converted to the solar constant by division by T_c. As an attempt to do this I use the solar-radiation measurements of the Commonwealth Observatory (Rimmer and Allen 1950, p. 40) which are reduced in such a way that the solar radiation at mean distance is given by

$$T_m \times 1{\cdot}940/1{\cdot}024.$$

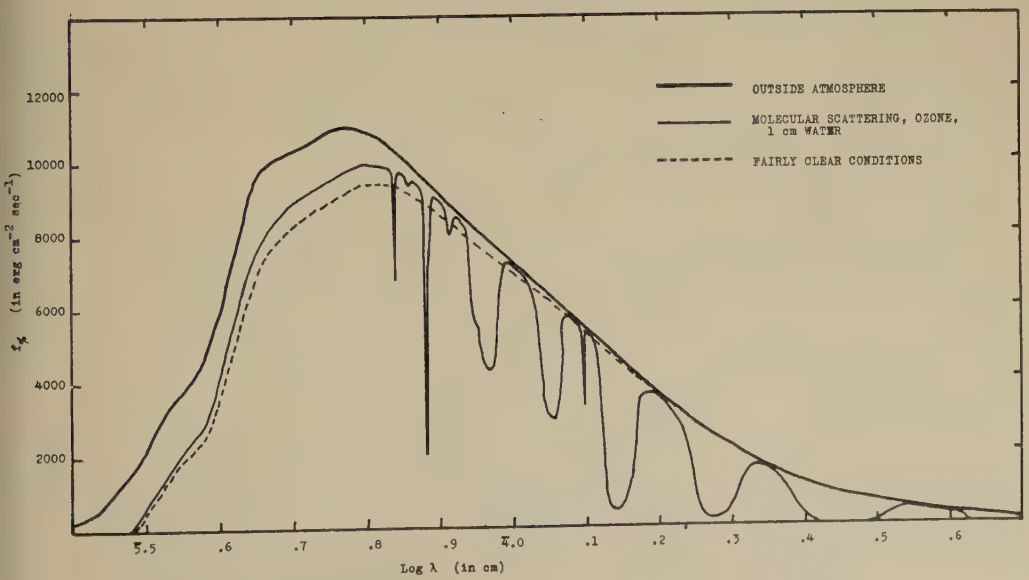

Figure 9. Effect of atmospheric absorption on the flux of solar radiation.

The value of T_m deduced for the conditions described was (0·793 + 0·0005) where the small correction represents dust absorption. The solar radiation was therefore 1·503 cal cm^{-2} min^{-1}, and the deduced solar constant 1·503/0·758

$$= 1{\cdot}983 \text{ cal cm}^{-2} \text{ min}^{-1}.$$

This simple procedure I have adopted for re-determining the solar constant has the advantage that it is almost independent of the procedures of the Smithsonian Institution. A recent discussion on the solar constant by Johnson (1957) leads to 2·00 ± 0·04 cal cm^{-2} min^{-1}. I would think the probable error not quite so high as this and would be willing to compromise with

$$1{\cdot}99 \pm 0{\cdot}02 \text{ cal cm}^{-2} \text{ min}^{-1}$$

as a suitable modern estimate of the solar constant. The corresponding effective solar temperature is 5,800°K.

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Absorption and emission in the atmospheric window from 770 to 1,250 cm^{-1}

By W. T. ROACH* and R. M. GOODY†
Imperial College, London

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SUMMARY

Combined absorption and emission spectra have been recorded in the atmospheric window from 770 to 1,250 cm^{-1} both in London and in the relatively clean air of Ascot, Berks. The absorption spectra have been partially analysed in terms of selective and continuous extinction. The continuous extinction outweighs the selective in both Ascot and London. There is good agreement between the observed selective absorption and existing theoretical and laboratory data upon gases known to be present in the atmosphere.

The opacity of the continuum shows an increase towards low frequency, and it is much greater in London than in Ascot; in both places it is positively correlated with water vapour. The average effective emission temperature of the continuum is not significantly different from that of water vapour.

The only hypothesis which seems to fit all the observations is that in London the continuum extinction is largely caused by aerosol extinction while in Ascot this is only of minor importance. This requires that the extinction at Ascot be caused by wings of distant lines; this old suggestion of Elsasser is re-examined and found to be plausible.

Attention is drawn to the important contribution which the continuum can make to the heat balance of the atmosphere below 2 km.

1. INTRODUCTION

The translucent region of the atmospheric spectrum between the water-vapour band at 1595 cm^{-1} and the carbon-dioxide band at 667 cm^{-1} embraces the maximum of the Planck emission curve for normal atmospheric temperatures, and through it the earth's surface and all layers of the atmosphere can radiate energy directly into space. Despite the importance of the region, however, quantitative measurements are few. Leaving aside for the present the 1,045 cm^{-1} ozone band, we know that there are many weak lines in the window, and the gases and transitions responsible for them have been identified in nearly every instance (Migeotte, Neven and Swensson 1956/57; Shaw, Oxholm and Claassen 1952). We know also from low-resolution spectra both of the sun and of artificial sources, that there is much absorption (Taylor and Yates 1957; Gebbie, Harding, Hilsum, Pryce and Roberts 1951; Adel and Lampland 1940) and emission (Sloan, Shaw and Williams 1956; Burch and Shaw 1957) in the region. Both are surprisingly large for selective effects alone, but the existing data are insufficient to decide whether a continuum exists as well, and if so what its cause may be.

In an attempt to resolve this uncertainty we undertook a series of simultaneous spectral observations of absorption and emission. High resolution is clearly desirable when there is the problem of separating selective and continuous extinction, but at the start of our observations the only available instrument was a conventional prism double monochromator. This has given rise to much unnecessary labour, and a certain lack of precision in the conclusions. Nevertheless, the indications are sufficiently decisive to warrant their discussion now, even though at a later stage, with new equipment, we hope for much more definite conclusions.

The 1,045 cm^{-1} ozone band has been mentioned, but we will not discuss it further since it has been the subject of a number of other papers by us (Goody and Roach 1956, 1958; Walshaw and Goody 1956) and by workers in the United States (Strong 1941; Watanabe 1943; Epstein, Osterberg and Adel 1956), and the position is, on the whole, reasonably satisfactory. The region from 1,100 to 948 cm^{-1} is therefore not considered in the following sections.

* Now with Meteorological Research Flight, Farnborough. † Now at Harvard University.

2 Observational methods

The spectrometer used for both absorption and emission measurements was a double monochromator with rocksalt prisms with an effective total basal length of 16 cm. The detector was a Golay cell and radiation passing the entrance slit was chopped at 12 c/sec.

If emission spectra of the sky and of two black bodies at different temperatures are recorded through the same optical system, the sky emission can be determined directly on an absolute scale (Goody 1957). To ensure that all the radiation passing the spectrometer slits comes from a black body of moderate size, the slit of the spectrometer was focused not at infinity but at a point outside the observing hut with a magnification of 2·6. For the emission measurements there was placed at the slit image a plane mirror, $5\frac{1}{2}$ cm × 7 cm, set at 45° to a horizontal axis about which it could rotate. By rotating the mirror the sky could be viewed at all air masses, either to the east or to the west, or the spectrometer could look down into Dewar flasks containing water or liquid nitrogen placed below the 45° mirror.

The recording of emission spectra proved to be particularly simple and trouble-free. The only difficulty initially was a zero drift caused by changes in the chopper temperature. This drift was minimized by giving the aluminium chopper a high polish, and by recording black-body spectra both before and after the sky spectra and making a linear interpolation to correct for small residual drifts.

A typical emission spectrum is shown in Fig. 1. Such spectra were usually recorded at air masses 1, 2, 3, 4, 5 and 6 with some repeats at opposite azimuths, during lulls in the rather more hectic solar observations. Unfortunately, on some occasions, by the time opportunity to take an emission run had occurred, cumulus convection had spoiled the sky near the horizon with the consequence that on 5 out of 14 days at Ascot absorption alone was measured.

Solar spectra had to be recorded over a wide range of air masses so that absolute values of the absorption could be found by extrapolation to the imaginary case of zero air mass. These observations were made with the 45° mirror removed thus exposing a siderostat mirror directly behind it. Since the position of the condenser mirror was not changed, the solar image was formed well in front of the entrance slit. A circular

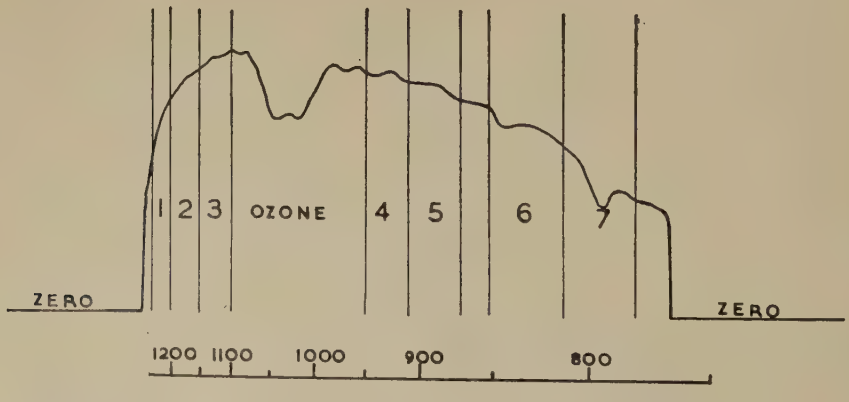

Figure 1. Emission spectrum recorded on 9 April 1956, with two air masses, 0·70 g cm^{-2} of water vapour in the zenith and with slit width 25 cm^{-1}. The vertical lines mark the boundaries of the spectral ranges which are discussed in the text. The spectrum is the difference between the emission from the spectrometer and the emission from the sky; the energy has the opposite sense to Fig. 2.

perture, just large enough to pass the solar image, and provided with cross-wires oated with magnesium oxide, was placed at the focus in order to locate the solar image vith respect to the optic axis of the spectrometer, and also to prevent the spectrometer rom looking at too much of the sky surrounding the sun. The maximum solar energy id not pass the spectrometer slits with this arrangement, and so they had to be opened o 7 cm^{-1} to obtain an adequate signal-to-noise ratio. A typical solar spectrum with this esolution is shown in Fig. 2.

The system had an important advantage. By using only the centre section of the iderostat mirror the aperture of the entire optical system was not influenced by the ngle of the mirror, and so remained constant through a series of observations.

For preliminary observations made in London (see below) the solar image was ormed on the entrance slits and the aperture was deliberately restricted to avoid diurnal hange. Simultaneous emission spectra were not observed. With the greater energy btainable slits could be closed to 5·5 cm^{-1}.

Two further corrections to the solar observations must be mentioned. Firstly, ve found from subsidiary measurements that the reflectivity of the siderostat mirror hanges by not more than 2 per cent over the range of angles of incidence used. No orrection was therefore made for this effect. Secondly, the gain of the amplifier varied hroughout the day, so an underrun and aged standard lamp was periodically swung n front of the condenser mirror, then by adjusting the spectrometer to the maximum f the lamp emission, near 5,500 cm^{-1}, a standardizing signal was obtained and correc- ions made to a standard gain.

Observations were first made from the roof of the Royal College of Science in outh Kensington, London, at an altitude of 35 m above sea level. Absorption and mission were very large and we soon concluded that we were measuring the properties f London pollution rather than those of the free atmosphere. Consequently the pparatus was transferred in February 1956 to the Imperial College Field Station t Ascot, Berkshire, 40 km west of London, at an altitude of 66 m above sea level, nd apparently free from direct London pollution even with light winds from that irection.

All observations reported in this paper were taken with the best sky conditions hat occur in this part of England.

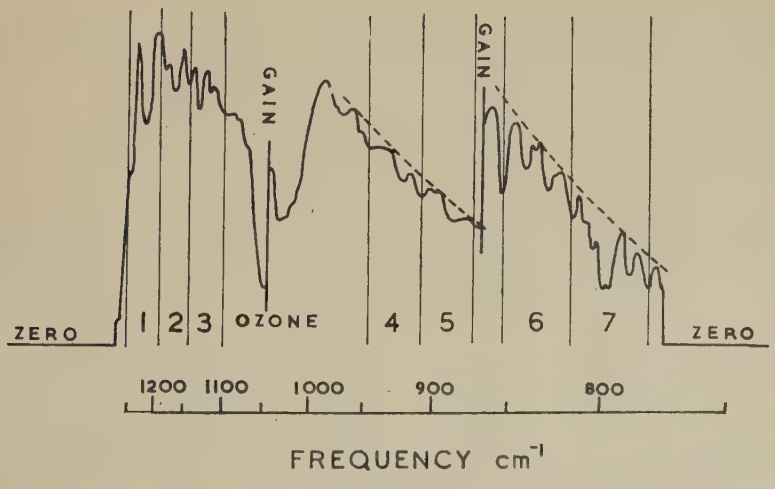

Figure 2. The solar spectrum recorded on 26 July 1956, with 1·20 air masses and 1·79 g cm^{-2} water vapour n the zenith. Slits 7 cm^{-1}. The vertical lines mark the boundaries of the spectral ranges. The dashed line s the background used for analysis of ranges 4, 5, 6 and 7. The gain of the amplifier has been changed by factor of two in the centre of the ozone band and also at about 860 cm^{-1}, in the gap left for this purpose.

3. Selective absorption

Solar absorption spectra were recorded on 14 days at Ascot and 6 days in London. The first step in the analysis was to divide up into spectral ranges, details of which are shown in Table 1.

TABLE 1. Details of spectral ranges

Range	Frequency (cm^{-1})	Width (cm^{-1})	Absorbing gases
1	1232-1198	34	H_2O (ν_2); CH_4
2	1198-1149	49	H_2O (ν_2); N_2O
3	1149-1100	49	H_2O (ν_2); N_2O
Ozone	1100-948	152	H_2O (ν_2 and rot.); CO_2; C
4	948-908	40	H_2O (rot.); CO_2
5	908-871	37	H_2O (rot.)
6	853-814	39	H_2O (rot.); CO_2
7	814-778	36	H_2O (rot.); CO_2

rot. = water-vapour rotation band; $\nu_2 = 1{,}595\ cm^{-1}$ fundamental vibration band of water vapour.

The second step was to separate selective from continuous extinction in ranges 4, 5, 6 and 7 by drawing a smooth curve through the seven points of maximum transmission, for inspection of a high-resolution spectrum reveals that for each of these points the separation between strong lines is at least 8 cm^{-1}. At these points therefore the spectrometer is sampling mainly radiation from the continuum despite the poor resolution.

The continuous extinction found in this way is very large. Normal limits for selective and continuous extinction are shown in Table 2 for ranges 4 and 7 at Ascot. For London the continuous extinction can be 2 to 3 times larger while the selective absorption remains the same.

TABLE 2. Approximate limits to selective and continuous extinction observed at Ascot (per cent)

	Air mass	Range 4	Range 7
Selective	1	2-5	15-25
	6	5-10	35-45
Continuous	1	5-15	10-25
	6	25-60	40-85
Total	1	5-20	20-40
	6	30-65	60-90

Selective absorption in ranges 4, 5, 6 and 7 is dominated by water vapour. For this gas a statistical model with an exponential distribution of line intensities (Goody 1952) gives a good representation of the mean absorption. This is a two-parameter model which can be defined in terms of

$$\eta = \frac{1}{\delta}(\pi\alpha\sigma\, m \sec z)^{\frac{1}{2}} \quad \text{and} \quad \xi = \frac{\pi\alpha}{\delta} \qquad . \quad . \quad . \quad (1$$

where σ = mean line intensity; δ = mean line spacing; α = mean line width; $\sec z$ = air mass; m = corrected zenith amount of water vapour†.

* The best mean line width to use in atmospheric problems is that suggested by Curtis (1952). This was evaluated separately for each day but in practice it varied little from day to day.

† Corrected for temperature in the usual way by scaling amounts according to the temperature dependence of σ, this being determined from theoretical line intensities of Benedict, Claassen and Shaw (1952). Water-vapour densities and air temperatures were taken from Meteorological Office ascents adjusted near ground level to screen temperatures and humidities measured on site

For absorption more than 5 per cent (as is normal in this work), the absorption becomes a function mainly of η, so theoretical values of ξ were used (see Table 3) making the absorption vary with η only. For each observed value of the transmission a value of η could be computed, and the regression of η on (m sec z)$^{\frac{1}{2}}$ then yields a value for $\frac{1}{\delta}(\pi\alpha\sigma)^{\frac{1}{2}}$. This method of treating the data has the advantage that they are all used together, and since the accuracy of an individual observation is poor, this is almost the only way to extract a significant result.

Allowance was made for the small amount of carbon-dioxide absorption in the following way. If we are dealing with absorption which is not very small, then for the statistical model

$$-\ln\tau = \eta \quad . \quad . \quad . \quad . \quad . \quad . \quad (2)$$

where τ is the transmission. For a regular model, which would be suitable for carbon-dioxide absorption, it is approximately true for absorptions which are not very small that

$$-\ln\tau = \text{const.}\,\eta \quad . \quad . \quad . \quad . \quad . \quad (3)$$

Since transmissions of different gases multiply it follows that

$$-\ln\tau \simeq \eta_w + \text{const.}\,\eta_c \quad . \quad . \quad . \quad . \quad (4)$$

where τ is the observed transmission and η_w and η_c are derived from the water vapour and carbon-dioxide transmissions respectively. We now assume that (4) is generally true and extract the two terms by a partial regression of ln τ on (sec z)$^{\frac{1}{2}}$ and (m sec z)$^{\frac{1}{2}}$, because carbon-dioxide differs from water vapour in that its concentration does not greatly vary from day to day.

The effect of carbon dioxide is not large. No contribution to the absorption could be detected in ranges 5 and 6. In range 4 there was a contribution of marginal significance, probably caused by P-branch lines of the 961 cm^{-1} band. In range 7 we found a significant correlation corresponding to 3 per cent absorption for one air mass and 10 per cent absorption for six air masses. This agrees well with data kindly supplied to us by Dr. L. D. Kaplan which indicated that there should be a mean absorption of about 2·3 per cent for one air mass and 9·6 per cent for six air masses.

TABLE 3. Observed and theoretical parameters for water vapour at 1,000 mb and 287°K

Range	$\frac{1}{\delta}(\pi\alpha\sigma)^{\frac{1}{2}}$				$\frac{\pi\alpha}{\delta}$
	Observed				
	London	Ascot	Migeotte	Benedict	Benedict
1	—	—	0·121	0·117	0·039
2	—	—	0·101	0·088	0·035
3	—	—	0·110	0·065	0·050
4	0·043 ± 0·008	0·031 ± 0·004	0·027	0·043	0·039
5	0·050 ± 0·013	0·050 ± 0·006	0·036	0·055	0·055
6	0·082 ± 0·024	0·060 ± 0·009	0·058	0·070	0·048
7	0·165 ± 0·051	0·154 ± 0·017	0·126	0·187	0·055

The results for water vapour are shown in Table 3. The observed results in columns and 3 were made with differing spectral resolutions. There is no significant difference, confirming that the resolution was sufficient to separate continuous and selective effects in this spectral range. Limits given in this and all other Tables are standard deviations.

Columns 5 and 6 in Table 3 are found from a best fit of the theoretical line strengt of Benedict, Claassen and Shaw (1952) to an exponential distribution. According to private communication from Mr. A. R. Curtis this is best performed by forming t expressions

$$\frac{\sigma}{\delta} = \frac{\sum_i k_i}{\Delta \nu} \quad . \quad . \quad . \quad . \quad . \quad (1)$$

$$\frac{\sigma}{\pi\alpha} = \frac{\left(\sum_i k_i\right)^2}{4\left(\sum_i (k_i\,\alpha_i)^{\frac{1}{2}}\right)^2} \quad . \quad . \quad . \quad . \quad (2)$$

where $\sum_i$ = sum over all lines in one spectral range; k_i = a line intensity; α_i = a lin width; $\Delta\nu$ = frequency width of the range.

The line widths used in this computation were calculated by Dr. L. D. Kapla for the rotation band, and for the ν_2 band the average of the line widths for the rotatio band was employed (0·055 cm^{-1} at STP).

The results attributed to Migeotte *et al.* in column 4 of Table 3 were obtaine by selecting isolated or near-isolated lines in their Atlas of the Solar Spectrum record on the Jungfraujoch, Switzerland, and evaluating line strengths using humidity an temperature data from Payerne (60 km from the Jungfraujoch), temperature dependen of line intensities from Benedict *et al.* (1952) and line widths computed by Dr. Kapla The few lines evaluated were compared with the theoretical values of Benedict *et a* and an average ratio was obtained for each range. Hence the results of column 4 Table 3 were derived from those of column 5.

For the rotation band, observation agrees well both with theory and the Jung fraujoch measurements. Theory should be reliable for the rotation band, but th confirmation is nevertheless welcome. From the point of view of this paper the agree ment between columns 2 and 4 shows that the method used to separate the continuu is satisfactory, and that the large extinction attributed to the continuum should be rea

Discrepancies between columns 4 and 5 could be caused by the combined effe of differences in the atmosphere above Payerne and the atmosphere above the Jung fraujoch and also to an incorrect choice of ν_2 band oscillator strength by Benedict *et a*

4. The continuum

In ranges 1, 2 and 3 no graphical separation of continuum and selective absorptio could be made and the data in Table 3, column 4 had to be used to estimate the lin absorption of water vapour. For range 1 the required data on methane are not availabl so we were obliged to abandon attempts to analyse this range. Absorption by th 1285 cm^{-1} band of nitrous oxide was computed from the data of Goody and Worme (1951) for ranges 2 and 3, and this together with the data on water vapour made possib an estimate of the continuous extinction.

Extinction by a continuum must follow Lambert's law, i.e., logarithm of transmissio proportional to air mass, while according to Eq. (2) selective absorption (if not weak gives proportionality between logarithm of transmission and the square root of ai mass. In Fig. 3 the logarithm of the deflection attributable to the continuum is plotte against air mass for one day of observation, showing clearly that Lambert's law i preferable to a square-root law of extinction.

We did not find this to be so when observations were first made in London; square-root law then seemed to provide a much better fit. This was a mysterio result, for even at that time we were convinced that we had to do with a continuum A possible explanation is that the material responsible for the extinction is not stratifie but forms a dome with width about six times its height. On a fine day such a dom over London is clearly visible from Ascot, and the explanation is therefore plausible In Table 4 extinction coefficients for London are evaluated for vertical incidence.

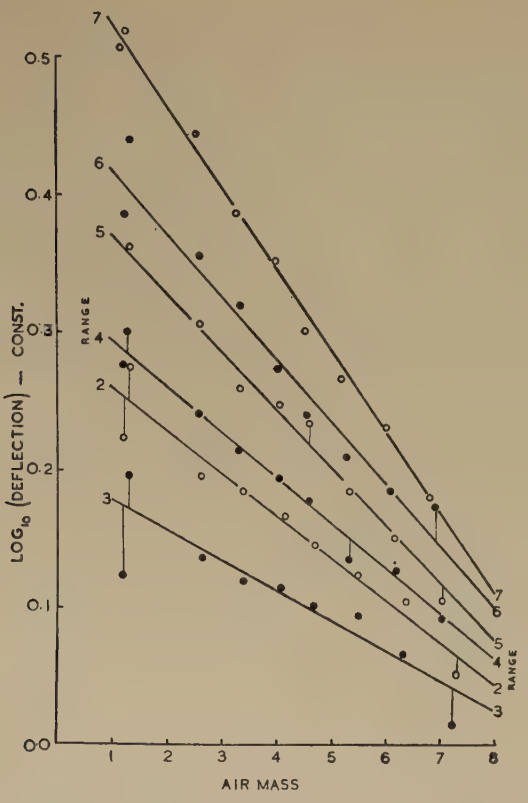

Figure 3. Logarithm of continuum deflection as a function of air mass. A different constant is subtracted for each range. 2 May 1956 at Ascot, 1·50 g cm^{-2} of water vapour in the zenith.

TABLE 4. CONTINUOUS EXTINCTION COEFFICIENTS, β, (air mass)$^{-1}$ × 10^3

Date	m (g cm^{-2})	2	3	4	5	6	7
				London 1955			
30 Aug.	2·07	230 ± 50	190 ± 50	500 ± 75	550 ± 75	650 ± 100	910 ± 150
17 Sept.	1·46	320 ± 100	180 ± 75	370 ± 50	380 ± 50	450 ± 75	400 ± 50
20 Sept.	1·74	300 ± 75	250 ± 75	220 ± 25	240 ± 25	300 ± 50	390 ± 50
29 Sept.	2·17	180 ± 50	160 ± 50	430 ± 75	400 ± 50	520 ± 75	670 ± 100
10 Oct.	1·10	180 ± 75	160 ± 75	330 ± 50	290 ± 50	360 ± 50	460 ± 75
31 Oct.	0·92	190 ± 50	180 ± 50	270 ± 50	310 ± 50	360 ± 50	500 ± 75
				Ascot 1956			
9 Apr.	0·61	50 ± 25	50 ± 25	65 ± 9	79 ± 8	75 ± 12	109 ± 13
18 Apr.	0·70	60 ± 25	50 ± 25	80 ± 12	88 ± 6	91 ± 6	143 ± 9
19 Apr.	0·66	60 ± 25	70 ± 25	56 ± 9	72 ± 14	60 ± 13	95 ± 20
20 Apr.	0·84	90 ± 25	70 ± 25	87 ± 4	95 ± 4	88 ± 4	132 ± 4
21 Apr.	0·84	80 ± 25	60 ± 25	80 ± 5	83 ± 7	86 ± 7	127 ± 5
23 Apr.	0·63	120 ± 25	130 ± 25	111 ± 6	129 ± 9	130 ± 7	174 ± 12
29 Apr.	0·80	90 ± 25	60 ± 25	72 ± 4	75 ± 4	95 ± 6	111 ± 6
30 Apr.	1·01	100 ± 25	80 ± 25	93 ± 5	110 ± 6	118 ± 4	161 ± 7
2 May	1·26	70 ± 25	60 ± 25	73 ± 6	91 ± 6	100 ± 9	137 ± 6
6 May	1·58	120 ± 25	80 ± 25	160 ± 9	173 ± 5	219 ± 6	284 ± 7
19 May	0·78	60 ± 25	40 ± 25	59 ± 1	70 ± 1	72 ± 1	104 ± 3
22 May	1·14	90 ± 25	80 ± 25	93 ± 10	95 ± 3	112 ± 3	157 ± 2
26 July	1·51	100 ± 25	60 ± 25	132 ± 9	149 ± 10	176 ± 8	189 ± 6
20 Aug.	1·64	110 ± 25	70 ± 25	143 ± 7	169 ± 10	207 ± 4	259 ± 6

The results of 20 days of observations are given in Table 4. There is clearly a large difference between London and Ascot and the results have therefore been analysed in two groups. In each group there is a relation between extinction and the total amount of water vapour in the zenith. Correlation coefficients and slopes of regression lines for ranges 4, 5, 6 and 7 are given in Table 5.

For ranges 4, 5, 6 and 7 at Ascot the standard deviation has been evaluated objectively. The indirect treatment of the continuum for ranges 2 and 3, and the 'dome effect' in London, involve the possibility of systematic errors; a rough estimate of these limits is given. For a continuum Curtis's treatment of pressure effects is identical to the more usual pressure-scaling approximation using the first power of the pressure. Water-vapour amounts in this section are scaled for pressure to 1,000 mb, but are not scaled for temperature, since this would be inappropriate for the wings of very strong lines, which vary little with temperature.

TABLE 5. CORRELATION COEFFICIENTS BETWEEN CONTINUOUS EXTINCTION COEFFICIENT (β) AND m, WITH SLOPES OF REGRESSION LINES

	Ascot (14 days)		London (6 days)	
Range	$R(\beta, m)$	$\Delta\beta/\Delta m$	$R(\beta, m)$	$\Delta\beta/\Delta m$
4	+ 0·86	+ 0·10 ± 0·03	+ 0·55	+ 0·23 ± 0·07
5	+ 0·80	+ 0·11 ± 0·04	+ 0·53	+ 0·29 ± 0·08
6	+ 0·86	+ 0·12 ± 0·03	+ 0·60	+ 0·29 ± 0·08
7	+ 0·82	+ 0·16 ± 0·04	+ 0·54	+ 0·37 ± 0·12

The correlation coefficients for Ascot shown in Table 5 are significant. An independent estimate of the variance attributable to experimental error indicates that this may

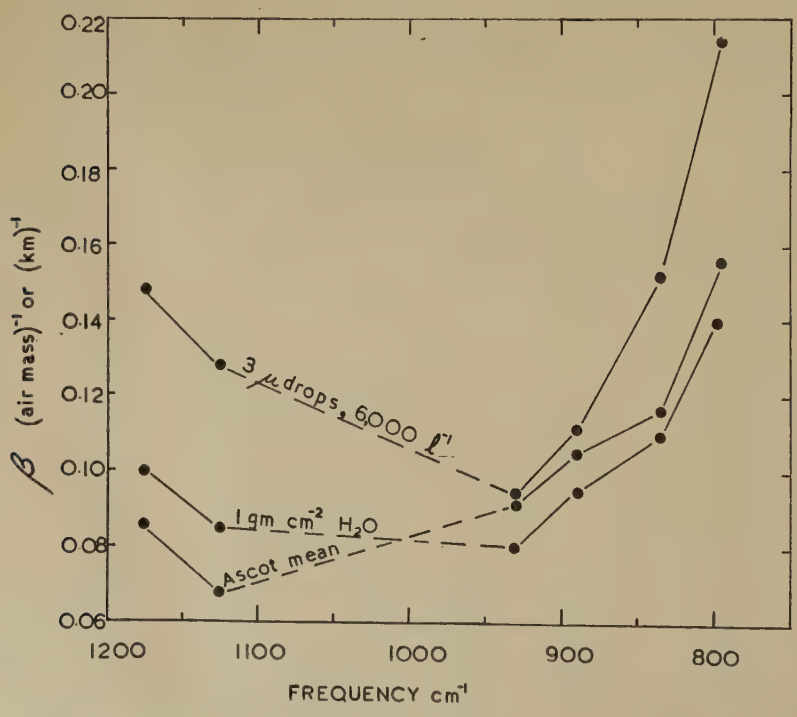

Figure 4. β as a function of frequency (see Tables 5, 7 and the Appendix).

account for all of the variance, in which case β and m may be perfectly related. This is not true for the London observations although the correlations are significant if the four ranges are taken together as estimates for the same population. Thus the correlations in the two groups differ, showing that a different mechanism operates on the two cases. This is confirmed by the systematic differences in the slopes of regression lines in the two groups.

An interesting feature of the continuum is that the extinction increases as frequency decreases from range 4 to range 7. This shows clearly in the average extinctions of Fig. 4.

Turning now to the emission of the continuum, Fig. 5 shows a typical set of emission measurements for one day at Ascot. The observational procedure was not fully developed for the London observations, and owing to difficulties with sky conditions, simultaneous absorption and emission observations were made on only 9 out of the 14 occasions at Ascot.

The emission is partly selective and partly continuous. If *either* selective and continuous optical depths are small *or* if both have approximately the same emission temperature, then the combined emission will be effectively the same whether the continuum overlies the selective emission or vice versa. Since both of these conditions are obeyed approximately, little error is involved in assuming arbitrarily that the selective layer lies below the continuum. From the data in Table 3 and the observed temperature and humidity profiles the absorption and emission of the selective component can be computed by standard means. This enabled us to compute the continuous emission, which, knowing the continuous extinction, we could express in terms of an emission

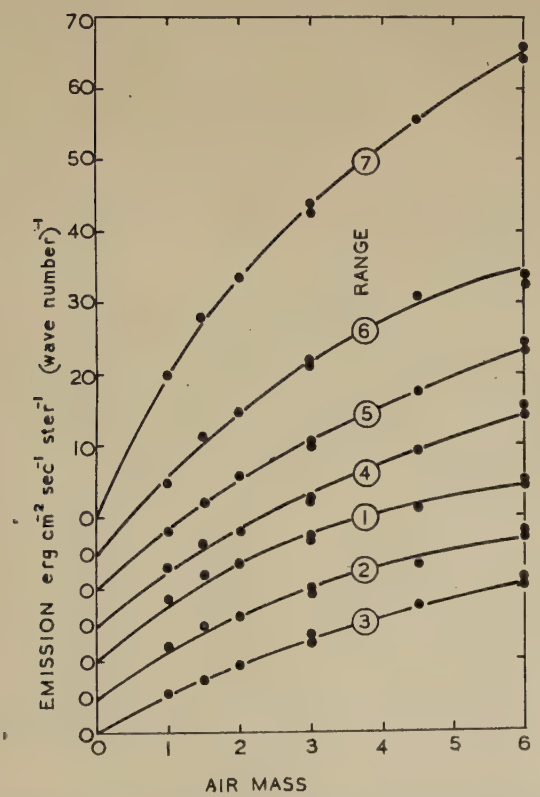

Figure 5. Emission as a function of air mass 21 April 1956. Zeros are offset. Repeated points are at opposite azimuths. Screen temperature = 276°K. 1·01 g cm^{-2} of water vapour in the zenith.

temperature T_c by application of Kirchoff's Law at each air mass. The difference $(T_s - T_c)$ is shown in Table 6 for ranges 4, 5, 6 and 7 where T_s is the emission temperature for the selective component. The mean difference is -2 ± 5°K.

TABLE 6. $(T_s - T_c)$ °K for 9 occasions at Ascot in 1956

Date	Screen* temp. (°K)	T_s* (°K)	$(T_s - T_c)$† °K Range 4	5	6	7
19 Apr.	277·5	270·0	+ 3	+ 8	− 6	− 14
20 Apr.	284·0	275·0	+ 7	+ 7	− 1	− 2
21 Apr.	283·0	274·0	+ 8	+ 5	+ 2	− 2
29 Apr.	283·0	276·0	− 3	− 5	− 2	− 7
2 May	287·0	278·5	− 5	0	− 5	− 7
6 May	295·0	286·5	− 1	− 1	− 2	+ 4
19 May	282·0	274·0	− 5	− 2	− 10	− 2
22 May	286·5	281·5	− 2	− 3	− 4	− 1
20 Aug.	288·0	282·0	− 7	− 7	− 6	− 7
		Mean	− 0·6	+ 0·2	− 3·8	− 4·2

Mean for all ranges $= -2 \pm 5$°K

* Rounded off to nearest 0·5°K. † Rounded off to nearest 1°K.

5. Discussion

As far as can be determined from these observations, selective absorption agrees as well with laboratory and theoretical data as can be expected. This leaves the nature and origin of the continuum as the outstanding feature of the observations requiring explanation.

Let us examine the two logical alternatives, namely that we are dealing with gaseous extinction on the one hand or particulate (solid or liquid) extinction on the other. If the extinction is to be caused by gases then it is almost certainly the residual effect of the far wings of distant strong lines. This suggestion was first explored by Elsasser (1938). If we are concerned with particulate extinction then it is convenient to examine it under two headings : either spherical water drops or solid particles of arbitrary shape and optical properties (dry aerosol particles). In London there is no reason for surprise if the extinction is partly caused by dry aerosol particles, but in the country a large contribution from this source is not to be expected. Volz (1957) working in various districts in Germany (some surburban) has shown that dry aerosol particles can only give rise to extinction coefficients near 1,000 cm^{-1} of about 2×10^{-3} km^{-1}. If the effective height of the haze layer is of the order of 1 km this will also be the extinction coefficient per air mass, and it is fifty times too small to explain our results.

Thus for Ascot, if aerosol scattering is to be important, it must be by water droplets, and only this hypothesis requires detailed examination. To judge by the results of Dessens (1949) and Wright (1939) it is not an unreasonable hypothesis in the humid atmosphere of southern England.

In the light of these considerations there seem to be only two acceptable interpretations of the results displayed in Tables 4 and 5 : *either* (i) the extinction at Ascot is largely caused by far wings of water-vapour lines, while the extinction in London is

additionally and principally caused by aerosol scattering, *or* (ii) the extinction at Ascot is principally caused by water droplets, while in London it is principally caused by dry aerosol particles.

One or other of these interpretations is required to explain why the Ascot extinction is the more closely correlated with the total amount of water vapour. The observed 'dome effect' of the London extinction does not help us to distinguish between the possibilities since both will account for it. The high correlation between Ascot extinction and amount of water vapour favours (i) but not definitely, since the size and scattering power of haze in this country is in some degree dependent on humidity.

In the Appendix we examine for self-consistency the hypothesis that the Ascot attenuation is caused by a haze of water droplets and we find serious difficulties which may be summarized as follows. To obtain a reasonable ratio of about three to one between extinction in the visible spectrum and near 900 cm^{-1}, it is necessary to suppose that droplets of 3 μ radius play an active role. The shape of the extinction curve for such droplets (see Fig. 4) has a qualitative resemblance to our observations, the rise towards low frequency being caused by increasing absorption. To obtain the right general level of extinction, however, 6,000 per litre of such droplets are required. According to an unpublished communication from Mr. W. G. Durbin, measurements from aircraft over southern England show that aerosol particles of radius greater than 2·5 μ are present in densities from 2 to 20 per litre in the lower layers of the atmosphere. This speaks strongly against the hypothesis. Moreover 3 μ droplets have a scattering albedo of about 0·33. It is shown in the Appendix that the results of Table 6 and the absence of any detectable discontinuity of emission on the horizon are inconsistent with such a high albedo.

The arguments in favour of hypothesis (i), although admittedly somewhat circuitous, seem to us to be substantial, but confirmation can be found in recent work by Arnulf, Bricard, Curé and Véret (1957). These workers have made extensive measurements of the extinction in all major windows in the spectrum from the visible to 1,000 cm^{-1}, in fogs, mists and hazes, the last category being distinguished by extinction coefficients less than 4·6 km^{-1}. Near 1,000 cm^{-1} fogs and mists have an extinction coefficient about half as great as in the visible spectrum. Hazes, however, have a ratio between one to ten and one to a hundred. The change is presumably attributable to a change in droplet size, smaller droplets giving rise to greater contrast between infra-red and visible extinction. We may expect that the clear-sky conditions of our work represent a further step away from fog and mist and aerosol scattering should give rise to even more contrast between infra-red and visible than is found for hazes. If so, our Ascot observations with a ratio of about one to three are not the result of aerosol extinction.

Finally, if we are to accept hypothesis (i) we must show that Elsasser's suggestion is plausible. We are concerned with displacements from line centres of 10^3 or 10^4 line widths, and a satisfactory theoretical treatment is, and will probably remain, lacking. *If* lines have the Lorentz shape the strong lines near band centres would give rise to the continuum shown in Table 7 and Fig. 4.

TABLE 7. CONTINUOUS ABSORPTION CAUSED BY WINGS OF LINES WITH A LORENTZ SHAPE

Range	2	3	4	5	6	7
β_{H_2O} (gm)$^{-1}$	0·10	0·085	0·07	0·075	0·08	0·09
β_{CO_2} (air mass)$^{-1}$	$< 10^{-3}$	$< 10^{-3}$	0·01	0·02	0·03	0·05
Total β	0·10	0·085	0·08	0·095	0·11	0·14

After Kaplan and Eggers (1956), Cowling (1950), Goody (1952) and water-vapour line widths computed by Dr. L. D. Kaplan.

The agreement between Table 7 and our observations, both as regards level and spectral form, is probably fortuitous, since the Lorentz shape is most unlikely to be valid at these displacements from a line centre. Benedict and Silverman (1954) have shown that for carbon dioxide and carbon monoxide resonant broadening the line wings show a deficit compared with the Lorentz theory, while for hydrochloric acid there is an excess. No data at all exist on line shapes of water vapour broadened by air, although some information for water-vapour resonant broadening is obtainable from the laboratory measurements of Benedict *et al.* (1952) who found no continuous absorption near 1,000 cm^{-1} for 0·5 g cm^{-2} of pure water vapour at 383°K (according to a private communication from Dr. Shaw, no absorption means less than 10 per cent). If we assume that line widths are proportional to the pressure and inversely proportional to the square root of the absolute temperature, a simple computation shows that the line width for collisions with air molecules must be at least half the resonant line width under the same conditions in order to reconcile our observations with these laboratory measurements. Becker and Autler (1946) find for the centre of a water-vapour line in the microwave region, that the non-resonant width is about one-fifth the resonant width. There is, of course, no reason for resonant and non-resonant shapes to be similar and no comparison in the wings can be inferred from a comparison of line centres. These data are therefore not incompatible and no direct evidence exists against Elsasser's hypothesis.

6. Conclusions

The agreement between theory and observation for the selective component in the spectral range from 770 to 1,250 cm^{-1} is as good as can be expected. Very much more precise measurements on a few lines in the rotation and ν_2 bands are now in progress with a spectral resolution better than 0·5 cm^{-1}.

As regards the continuum there seems little doubt that in London the aerosol contribution is predominant, but in the country the evidence points the other way, supporting Elsasser's proposal that far wings of strong lines may be important. This suggestion could be tested in the laboratory, given an adequate absorption path-length. Alternatively sufficiently accurate measurements of selective and continuum attenuation in the solar spectrum could show that they are too closely related for them to have anything but a common cause.

We are attempting to follow the second course with recently-constructed equipment. A large grating spectrograph, suitable for emission studies has been built (Goody 1957), with which a resolution of 1 cm^{-1} can be reached. Our existing spectrometer has been equipped with gratings and an indium-antimonide filter to separate orders, giving a resolution of 0·5 cm^{-1} or better in solar spectra. A high-temperature black body has also been constructed with which we hope to measure the solar intensity in some of the windows near 1,000 cm^{-1}. When this has been done continuum extinction will be obtainable from spot readings, and if it should appear that the continuum is not caused by line wings we should be able to make sufficient observations to study its meteorology and hence determine its origin.

Apart from the general interest attaching to an understanding of the continuum, it is also of some thermodynamic importance. Möller (1942) has shown that an absorbing layer of similar optical thickness to that which we observe can, if restricted to the lowest one or two km, give rise to a cooling of about 1°K day^{-1}, i.e., about the same as the cooling caused by selective absorption at all wavelengths. The continuum may therefore be a significant factor in determining the planetary dynamics.

Acknowledgments

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Appendix

Extinction by water drops near 1,000 cm^{-1}

According to Aden and Kerker (1951) and Güttler (1952), a liquid-covered nucleus behaves optically like the pure liquid. Atlas, Kerker and Hitschfeld (1951) have shown that small departures from a spherical shape do not greatly influence the optical properties of droplets. Computations for spherical water drops (van der Hulst 1957), and particularly those of Johnson and Terrell (1955), can therefore be used confidently for the non-ideal circumstances occurring in the atmosphere.

Measurements of extinction were not made in the visible spectrum, but according to Wright (1939) a mean coefficient for fine days of about 0·33 km^{-1} might be expected, while for Ascot the extinction coefficient at 900 cm^{-1} is about 0·1 km^{-1}, if we assume an equivalent height for the haze of 1 km.

The meaning of this difference between visible and infra-red extinction has been considered in terms of what is known about the size distribution of aerosol particles. This leads however to a rather complex discussion and the same general conclusions

are reached by considering a single size of particle. Obviously this is a fiction, but it is reasonable to suppose that the combination of size distribution and optical properties may give preponderant importance to a narrow range of particle sizes.

Following Johnson and Terrell (1955) and assuming average optical properties for ranges 4, 5 and 6, it may be shown that droplets of 3 μ radius are required to give the observed contrast between visible and infra-red. Smaller particles give too large contrast and larger particles too small. 3 μ droplets have a scattering cross-section in the visible spectrum twice the geometric cross-section, and hence we may compute that to give an extinction coefficient of 0·33 km^{-1} we require 6,000 droplets per litre. As remarked in the text, this is an improbable figure.

The scattering albedo of these 3 μ drops will be about 0·33 (van de Hulst 1957, p. 180), that is to say, one quantum is scattered for each two that are absorbed. The scattering is mainly forward and, from the same source (p. 153), we estimate that the total forward scatter is four times the total backward scatter.

Now consider an optically thin, isothermal layer of droplets. The emission at a zenith angle z is equal to the opacity multiplied by the source function, $J(z)$ (see Kourganoff 1952). By definition

$$J(z) = (1 - a)\,B + a\,I(z) \quad . \quad . \quad . \quad . \quad (7)$$

where a = the scattering albedo

B = the Planck function for the layer

I = the effective mean intensity having regard to the scattering phase function of the medium.

For vertical viewing

$$I(0) = 1/5\,B \quad . \quad . \quad . \quad . \quad . \quad (8)$$

because 4/5 of the scattering is away from the earth's surface, while for horizontal viewing half of the scattering is from the earth's surface and hence,

$$I(\pi/2) = 1/2\,B \quad . \quad . \quad . \quad . \quad . \quad (9)$$

Thus

$$J(0) = B\,(1 - 4/5\,a) \quad . \quad . \quad . \quad . \quad (10)$$

$$J(\pi/2) = B\,(1 - 1/2\,a) \quad . \quad . \quad . \quad . \quad (11)$$

According to (10) and (11), excepting the case of pure absorption ($a = 0$), the source function varies with zenith angle. Using our new grating spectrometer in the 900 cm^{-1} window we have made very accurate observations to try to see whether the source function departs from constancy. Unfortunately with emission measurements alone the optical thickness of the continuum is not known and it proves to be very difficult to obtain a discriminating answer.

Given combined absorption and emission measurements, as for the 9 cases described in this paper, it is possible to discover whether the source function falls below the Planck function. This is in effect what was done to obtain figures in Table 6, and a difference $T_s - T_c = +1$°K corresponds to a source function 1·6 per cent smaller than the Planck function. This suggests that the albedo is probably less than 0·1.

The horizon from our observatory at Ascot is formed by trees rising 2° above the true horizon at a distance of about 1 km. Following our simple model a little further it is to be expected that, when looking at the trees, the source function will be simply B. From Eq. (11), therefore, there should be a discontinuity of ($\frac{1}{2}\,aB$) in the emission at the upper boundary of the trees. Our new grating spectrometer was stopped down until it embraced only 1° in the vertical and a search was made for this discontinuity. A set of results is shown in Fig. 6. There is clearly no evidence for a discontinuity, and although some uncertainties exist owing to the importance of the precise temperature structure of the lower layers of the atmosphere, it is thought that a discontinuity of 0·05 B would be obvious. This suggests that the albedo is less than 0·1.

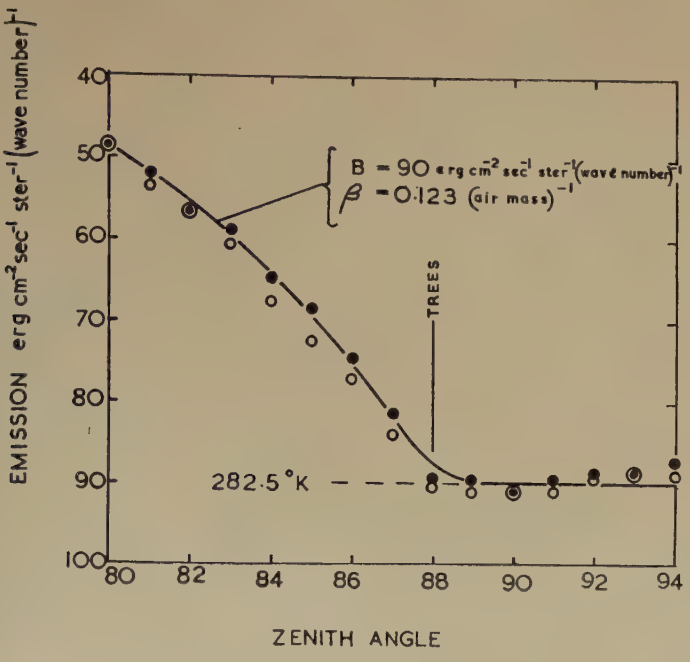

Figure 6. Emission near the horizon in the 900 cm^{-1} window. Elevation increasing (o), elevation decreasing (●). 27 Sept. 1957. The full line gives the emission of a purely absorbing layer of optical thickness 0·123 at a temperature of 282·5°K. Spectrometer aperture 1°.

Unofficial reports have circulated to the effect that the sun has an aureole in the 900 cm^{-1} region of the spectrum. If this were true it would be direct evidence for aerosol scattering. Careful measurements were made at 0·58° and 2·5° from the centre of the solar disc, and the emission was found to be the same within 0·01 B. However, 3 μ droplets would only give rise to a difference of 10^{-5} B for an optical depth along the solar beam of 0·2, so that our failure to detect the aureole is not surprising. It can be asserted with confidence that an aureole at 900 cm^{-1} is not detectable by any conventional technique.

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Some observations on flying locusts and atmospheric turbulence in eastern Africa

By R. C. RAINEY

Desert Locust Survey, East Africa High Commission

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SUMMARY

Flying swarms have been observed to vary in vertical extent from a few metres to several thousand metres, with the spacing of the locusts in them ranging from more than 10 locusts/m^3 down to the order of 0·001/m^3. Some characteristic effects seen, from ground and air, are described in qualitative terms.

Two well-documented records of flying locusts in numbers at some 2,000 m above ground were both associated with lapse rates close to the dry adiabatic, from the ground up to at least the level of these locusts. On two other occasions detailed observations of vertical temperature distribution have been made in the immediate vicinity of large swarms flying up to 1,100-1,700 m above ground, and in both these cases the topmost locusts were within 150 m of the upper limit of superadiabatic or adiabatic lapse rates from the surface. Comparable observations on a swarm in which all flying locusts were below 6 m demonstrated isothermal conditions, with the air temperature at 50 m within ½°F of that at the surface.

All adequately-documented records of swarm displacement so far available have been directly down-wind at ground speeds which for the larger swarms studied have approximated to the speed of the corresponding wind. The smaller swarms, most of them less than 10 km^2 in extent, have shown lower flying heights, relative to the corresponding vertical extent of dry adiabatic lapse rates, together with ground speeds which were also low, in relation to the wind speeds concerned and to the corresponding performance of the larger swarms.

There is some evidence of an association between rain and low flying heights; and consideration of heat exchange at the ground surface indicates that the energy normally available for convection, in the arid regions frequented by swarms, is likely to be drastically reduced after rain.

Quantitative evidence on the strength and distribution of the vertical components of air movement likely to be encountered by flying locusts has been provided by pilot-balloon and accelerometer data. Twin-theodolite pilot-balloon observations in central Somalia have shown that during much of the day some 10 per cent of uniformly-distributed locusts in flight at about 20 m at any one time could be expected to experience up-currents exceeding their gliding sinking-speed. Gust-spectra recorded by aircraft accelerometers within and around high-flying swarms have demonstrated up-gusts, exceeding a value equivalent to the sinking-speed of a gliding locust, at about fifty points per km^2 among the higher-flying locusts at any one time.

The possible use of locusts as indicators of air movements is discussed, with a summary of the evidence available on the contributions made by the active behaviour of the locusts themselves to the effects observed. Attention is directed to the importance of gregarious behaviour in the continued cohesion of individual swarms, observed over many days and hundreds of kilometres, despite the disruptive effects both of atmospheric turbulence and of the apparently random orientation of the flying locusts themselves. The order of magnitude of these potentially disruptive effects is estimated for a particular case, and shown to be large compared with the variations actually observed in the extent of the swarm concerned.

1. INTRODUCTION

Locust swarms in flight exhibit a protean variability of structure, ranging from that of stratiform swarms sometimes only a few metres in vertical extent, with volume densities of the order of 1 to 10 flying locusts/m^3, to that of cumuliform swarms sometimes several thousand metres in vertical extent, with volume densities of the order of 0·001 to 0·1 locusts/m^3. Such differences in formation are often exhibited at different times by one and the same swarm, but have not been found to show any obvious, consistent association with simple environmental factors such as air temperature (Rainey and Waloff 1951; cf. also Figs. 1 and 2 (Plates IV and V)). These differences in swarm formation appear to have a marked effect on the efficiency of aircraft spraying operations against flying swarms. Thus field assessments of such operations against swarms flying at high density in stratiform formation have demonstrated locust kills, per unit of insecticide applied, which were greater by an order of magnitude than those given by roughly comparable operations against swarms in cumuliform formation at low volume densities (Rainey 1956, 1958).

Figure 1. A small stratiform swarm, from the air. Swarm covering 1 km^2, with locusts flying mainly below 20 m, at air temperature of about 35°C at ground level. Wajir district, Northern Frontier Province, Kenya; 1715, 13 Jan. 1953. (Photograph by H. J. Sayer).

Figure 2. Part of a large cumuliform swarm. Swarm covering 75 km^2; seen from 700 m above ground (still below level of topmost locusts), looking up-sun; only part of swarm visible is that in direction appropriate for specular reflection. Air temperature at ground level 27°C. Near Hargeisa, Somaliland; 1545, 2 Oct. 1953. (Photograph by H. J. Sayer).

Regnier (1931) described locust swarms seen during the winter months in the High Atlas, with columns of locusts rising to great heights in fine weather, and first suggested that the heights at which locusts fly might be affected by convection currents. Field observations on flying swarms in recent years, with records of the associated meteorological factors, have provided evidence, summarized in the present paper, that the interpretation of the observed variations of swarm structure does in fact involve not only the behaviour of the flying locusts but also the degree and nature of convective turbulence in the air in which they fly.

Except where otherwise indicated, the observations are first-hand, and relate to swarms of sexually-immature adult Desert Locusts (*Schistocerca gregaria* Forskål). Most of the observations were secured in the course of research work on swarm behaviour and control development, reviewed elsewhere (Rainey, Waloff and Sayer 1955, Rainey 1958) and carried out in co-operation with Miss Z. Waloff of the Anti-Locust Research Centre, and Mr. H. J. Sayer of the Desert Locust Survey, to whom I am greatly indebted for many useful discussions as well as for the use of unpublished material (indicated by initials). The present paper is based on one prepared at the suggestion of Dr. R. S. Scorer for the 5th Congress of the Organisation Scientifique et Technique Internationale du Vol à Voile, at Buxton in 1954, and presented on my behalf by Mr. H. J. Sayer at the symposium of 24 July on ' convection and soaring in nature.'

2. Qualitative observations on swarm structure

The structures exhibited by flying swarms are so variable, often so complex, and in detail so little understood that any attempt to fit the limited number of observations yet available into a generalised description would be necessarily inadequate and probably misleading. In the following account, particular observations are accordingly quoted in detail, to illustrate the range of effects which have so far been seen.

In recording observations on flying swarms it has been found convenient, as already indicated, to distinguish between stratiform swarm structure, of flattened shape and limited vertical extent, and cumuliform structure, billowing or towering over a substantial vertical extent. Only exceptionally have high-flying swarms been seen apparently clear of the ground, with a space beneath almost empty of locusts – on two occasions, recorded photographically, in Somaliland in 1952 (Z.W.), at Burao at 1715 (LMT – used throughout) on 10 Sept. and near Hargeisa at 1440 on 28 Sept., and twice in the experience of the writer, the first time concerning a swarm covering 2 × 2 km near Korogwe in Tanganyika at 1530 on 27 Feb. 1954, when an aircraft passing beneath the swarm encountered only thinly-flying locusts, and the second occasion relating to a swarm 12 × 12 km and some 600 m above the ground, without apparent ' roots,' near Mandera in Somaliland at 1640 on 6 Sept. 1957. All four of these swarms exhibited a relatively compact, somewhat lenticular form, and were flying in the immediate vicinity of active cumulonimbus centres.

Swarms recorded as stratiform (Fig. 1, Plate IV) have ranged from a few metres to a few tens of metres in vertical extent, with the flying locusts constituting a sheet which, in all cases so far studied in detail, was closely applied to the surface being traversed; the sheet was sometimes deformed, and occasionally perforated, by prominent features of the terrain. In most cases the flight of the individual locusts was intermittent, with considerable numbers temporarily settled beneath those flying at any one time, causing the swarm to progress in a rolling manner. Apparently continuous flight, by all the locusts in a stratiform swarm, has been seen on a few occasions, at relatively low air temperatures (20°C to 27°C; Waloff 1953).

Locusts in stratiform swarms have often been seen to fly in sufficiently close formation to obscure objects at distances of less than a hundred metres within the swarm, and at times (e.g., 1600, 18 Feb. 1954, in the Usambara mountains near Lushoto, Tanganyika, following some disturbance by spraying aircraft) have largely concealed the ground when such swarms were viewed from above; volume densities up to 14 locusts/m^3 (Rumuruti, Kenya Highlands, 1945, in a swarm of about 10 m deep – Gunn, Perry, Seymour, Telford,

Wright and Yeo 1948) have been recorded photographically in such conditions. Wh stratiform swarms have been viewed from the air, at a distance and at an appropria angle with respect to the sun, an intense and unmistakable glitter has been observe resulting from specular reflection from the wings of the whole swarm.

Swarms recorded as cumuliform have ranged in vertical extent from about a hundr to more than a thousand metres, with volume densities generally lower, by several orde of magnitude, than in stratiform swarms, and characteristically highly variable fro point to point within the swarm. Thus numerous aircraft traverses through such swarm have given an impression of series of brief periods during which flying locusts were bei struck by the aircraft, alternating with longer periods without them, representing ga or spaces within the swarm.

The detailed structure of cumuliform swarms still challenges even qualitative descri tion. From the air, at a short distance outside and above such a swarm, and in a directio appropriate for specular reflection, a very characteristic reticulated pattern, with more less regularly spaced, wispy individual elements, often forked or Y-shaped, has bee seen on a number of occasions (e.g., 1800, 7 Mar. 1955, in a swarm covering about 90 km over the foothills of the South Pare mountains, Tanganyika). At other times the gener effect has been of shining, gauzy veils or columns, appearing, moving and disappearin with the changing angle of viewing (e.g., Fig. 2, Plate V). On another occasion it w noted that, with the passage of the observing aircraft, a point of intense specular reflectio travelled up such a column, along perhaps a quarter of an apparently spir path. The impression given by both types of effect – reticulation and veil – is evanescer and ethereal, but objective evidence of both has been secured by colour cinephotography on the dates quoted; attempts at black-and-white photography of these effects hav so far been less satisfactory.

Viewed from a lower altitude, against the sky, cumuliform swarms often resembl smoke, and from a distance have at times been mistaken for bush fires, but have usuall been distinguishable by a characteristic fibrous or streaky texture, with groups of hig flying locusts sometimes giving the impression of clouds, streaks, or wisps relativel separate from the main swarm. K. F. Sawyer has pointed out that this fibrous streakines is paralleled in appearance (under admittedly very different air-flow conditions) by th ' wispiness ' of mist freshly forming in a valley, and by the striae in a cloud of tobacc smoke drifting in the still air of a closed room.

On occasion an effect as bold as a charcoal stroke has been seen against the sky sloping downwards at a marked angle towards the horizon, and subsequently fading within a few minutes as the aircraft continued to approach the swarm. Another effect, lasting for several seconds and seen on two occasions while flying past large swarms at a distance of several kilometres and at a height below that of the topmost locusts, has resemble a thin plume of smoke from a moving train on a calm day, appearing to travel along the swarm in the same direction as the aircraft (1505/4 and 1545/21 Aug. 1957, Hargeisa Borama area, Somaliland).

The top of a cumuliform swarm, seen from an aircraft flying close abreast of it, has sometimes been represented only by a thin wisp of locusts (e.g., p. 338), but on other occasions (e.g. Feb. 1952, Wajir, Kenya) has shown a particularly clear-cut boundary, giving the impression of a distinct skin of flying locusts. Evidence that relatively high volume densities can at times be encountered well clear of the ground has also been provided by photographs taken near the top of a cumuliform swarm, in which individual locusts could be counted against the background of part of a second aircraft flying in company (1050, 11 Feb. 1952, near Wajir, Kenya), and gave approximate estimates of 0·4 to 0·7 locusts/m^3 – considerably higher than most of the values recorded for the lower levels of such swarms by photographs from the ground.

Observations from the ground within such swarms have recorded, in all adequately documented cases, substantial numbers of temporarily-settled locusts beneath the fliers, sometimes with indications of a marked association between gusts of wind and temporary apparent increases in the number of locusts seen flying low and settling. Kennedy (1951)

has similarly recorded that ' whenever a low-flying swarm was observed in windy weather, ... the proportion of insects flying . . . . and its height of flight decreased when the wind rose and increased when it fell.'

Locusts in cumuliform swarms have been repeatedly observed to glide (p. 343), a form of flight behaviour sometimes exhibited simultaneously by the majority of airborne locusts in sight over short periods, and at times seen also in stratiform swarms. Another effect which has often been seen on looking upwards through such swarms, is a swirling, circular or spiral movement, affecting numbers of flying locusts at the same time within a limited area of the swarm, and with the individual locusts moving at speeds (both linear and angular) which appeared large compared with the performance to be expected of the locusts themselves in still air. This effect also has been recorded by cinephotography, in Somaliland in 1952 (mature swarm near Las Anod, 23 Sept. 1952; Z.W.).

In observations from the ground at a distance outside cumuliform swarms, Gunn *et al.* (1948) recorded that on some occasions, in Kenya in 1945, ' great pillars (rose) from the general line of the swarm, looking . . . . like peculiar smoke.' Another striking example of this effect was observed in a broad valley among the eastern foothills of Mount Kenya in September 1946. ' Observations were made from a point . . . . overlooking the valley from the west and about 5 miles away from the swarm, which was moving down-wind in a . . . . north-westerly direction. As it progressed over the valley, smoke-like pillars, slightly inclined in a northerly direction, with bands of denser concentration appearing and disappearing within them, were seen to rise above the general level of the swarm and to increase rapidly in height to some 1500 ft above the valley floor. Each pillar retained its roughly columnar form for a time and then disintegrated and finally disappeared from view; while some were disintegrating others were developing, so that at times several columns could be seen at once, progressing down-wind in the same direction as the main swarm. Vigorous convection was . . . . demonstrated by . . . . dust devils rising to great heights from the valley floor. . . . . Measurements on photographs indicate that the distance between successive locust pillars was of the order of a mile ' (Rainey and Waloff 1951).

Gunn *et al.* (1948) describe how, in the pillars already mentioned, ' lines of greater density appear and move more quickly than individual locusts could possibly do, giving the impression of violent surging and even gyration of the whole pillar.' A ciné sequence, taken near Mtito Andei in Kenya in February 1955, clearly records swirling lines appearing and disappearing, inclined at varying angles, passing rapidly across the field of view, mainly without obvious connection with the ground, and at times apparently branched, giving temporarily suggestions of a reticulated pattern. These swirling lines, of higher apparent density of flying locusts, were seen, at distances beyond those at which individual locusts could be distinguished, by looking horizontally into a passing swarm from a point immediately outside it; and the effect differs both in scale, and perhaps in kind, from the circular or spiral swirling movement of individual flying locusts already mentioned as seen by looking vertically upwards from within such a swarm.

In view of the evidence of considerable gaps or spaces in the upper parts of such swarms, with flying locusts encountered at intervals at relatively high densities, it is suggested that these lines of apparently higher density, appearing and disappearing and often in rapid apparent movement, sometimes clearly associated with the relative movement of the swarm and the observer, may be interpreted as tangential views of sheets of flying locusts, often curved and too thin to be seen from a distance when viewed from a direction normal to the sheet. H. J. Sayer has pointed out that the reticulated pattern might be similarly interpreted, as evidence of a cellular or foam-like structure, with the forked elements of the pattern representing corners of cells or points of contact of bubbles.

Much of the adult life of the locust (which begins with the acquisition of effective wings at the final moult) is spent in a state of sexual immaturity, in which most of the striking long-range flights are made. Swarms of sexually mature adults have nevertheless exhibited a range of structure similar to that already described for immature swarms. A mature stratiform swarm, for example, some 3 km^2 in extent, was seen from the air

at 1540 on 12 March 1954, near Narok at an altitude of 1,800 m in the Kenya Highlands at an air temperature of about 29°C, with the characteristic yellow colour of the mature locusts in conspicuous contrast with the reddish, immature swarms seen earlier that season. An example of a large mature cumuliform swarm, covering an area of some 40 km^2, was seen from the air at 1300 on 20 June 1953, near Adi Ugri at an altitude of 2,000 m on the Eritrean plateau, with an air temperature of 26°C at ground level and flying locusts up to 450 m above it; in the same swarm at 1700 the previous afternoon (H.J.S.), flying locusts had been observed up to 650 m above the ground (i.e., 2,650 m above sea-level), showing the characteristic reticulated pattern already described (p. 336). Despite this flight activity, the majority of the females, in a sample from this swarm examined by Dr. W. J. Stower, had fully-developed eggs.

3. Vertical temperature distribution and height of flight

Circumstantial evidence of the part played by convection currents in the development of some of the structures assumed by flying swarms has been provided by the associated vertical gradients of air temperature. Thus during observations in Wajir district, Kenya, in February 1952, the most extreme example of stratiform swarm structure seen (some 5 min before sunset on 13 Feb.), with all flying locusts below about 6 m, was associated with isothermal conditions, with the air temperature at 50 m within 0·3°C (½°F) of that at the surface (and, at 34°C, incidentally well within the range necessary for sustained flight – Waloff and Rainey 1951, Weis-Fogh 1956); while on seven occasions during this same week air temperatures were recorded at the upper levels of cumuliform swarms over 300 m in vertical extent, together with corresponding temperatures near the ground, and in every case a superadiabatic lapse rate of 4½°C to 5½°C in 300 m was found.

Evidence of a close association, under dry, subtropical conditions, between vertical temperature distribution and the upper limit of convection currents from the surface has been provided by gliding experience in the Transvaal; thus, for example, the maximum height attainable by sailplane in each of five successive convective up-currents, during one particular fair-weather flight in the Pretoria area, increased progressively from 490 m above starting level at 1028 to 1280 m at 1255, and each of these five individual altitudes was within 30-120 m of the height of intersection of the path curve, through the corresponding surface temperature at the time, with the environment curve recorded during an aeroplane ascent earlier the same morning (Rainey 1947).

The vertical temperature distribution up to above the top of large cumuliform swarms in fair weather over relatively flat country, has been determined on two occasions, and in both cases the topmost locusts were within 150 m of the base of an isothermal layer marking the upper limit of superadiabatic or adiabatic lapse rates from the surface. Thus Fig. 3 shows the temperature distribution recorded in the course of aircraft observations on a large swarm during the NE monsoon in the Northern Frontier Province of Kenya. The topmost locusts were a thin wisp seen some 150 m above the recorded base of the isothermal layer, at a height which corresponds, as indicated, with the degree of penetration of the stable layer which would be expected of air rising from the surface with an initial temperature of 31°C; this temperature was actually attained in the screen at Wajir at 1145, while these aircraft observations were in progress. Again, in Fig. 4, aircraft observations by F. W. Bird, made during the SW monsoon in Somaliland, showed a lapse rate slightly exceeding the dry adiabatic from ground level to a height of 1,200 m, which was some 130 m above the topmost recorded locusts. Above 1,200 m was an isothermal layer, at a temperature of 18°C, extending to at least 1,500 m (Z.W.).

Well-documented reports of flying locusts at greater heights have been provided by British Overseas Airways Corporation aircraft on two occasions in eastern Africa (Rainey and Waloff 1951). At 1155 on 10 Aug. 1946 some 50-100 locusts were observed at 2,100 m from a Lodestar aircraft near Bulhar on the Somaliland coast, together with fair-weather cumulus with bases at 2,700-3,000 m. A true air temperature of 20°C was recorded

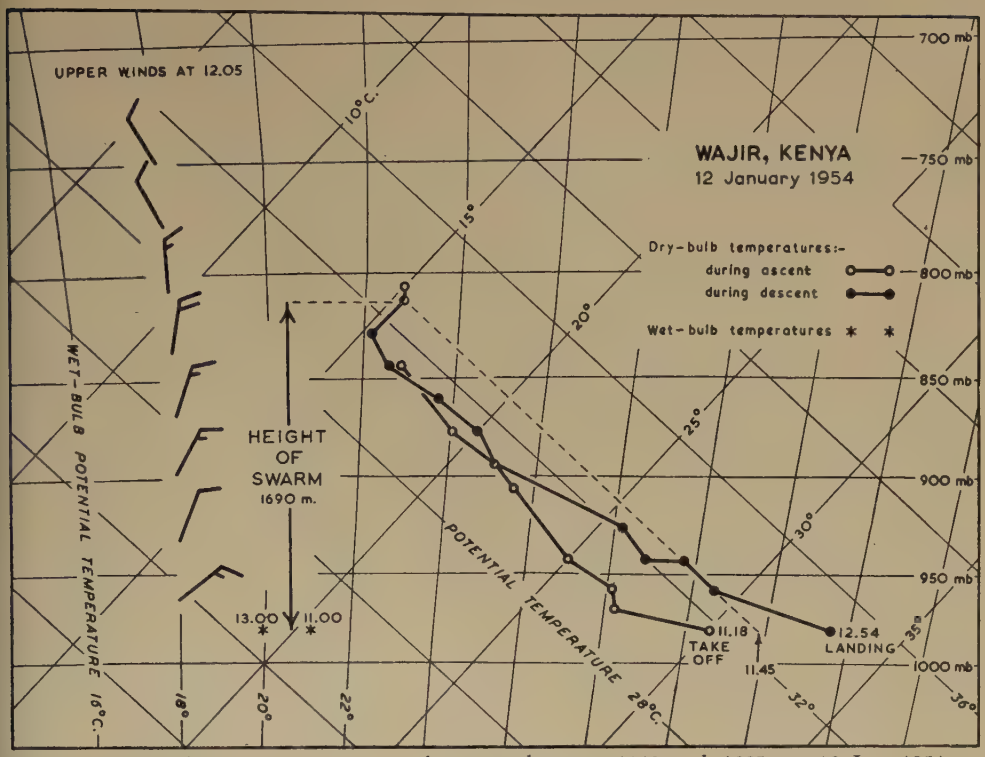

Figure 3. Observations on swarm 24 × 6 km, seen between 1140 and 1225 on 12 Jan. 1954, some 70 km ESE of Wajir.

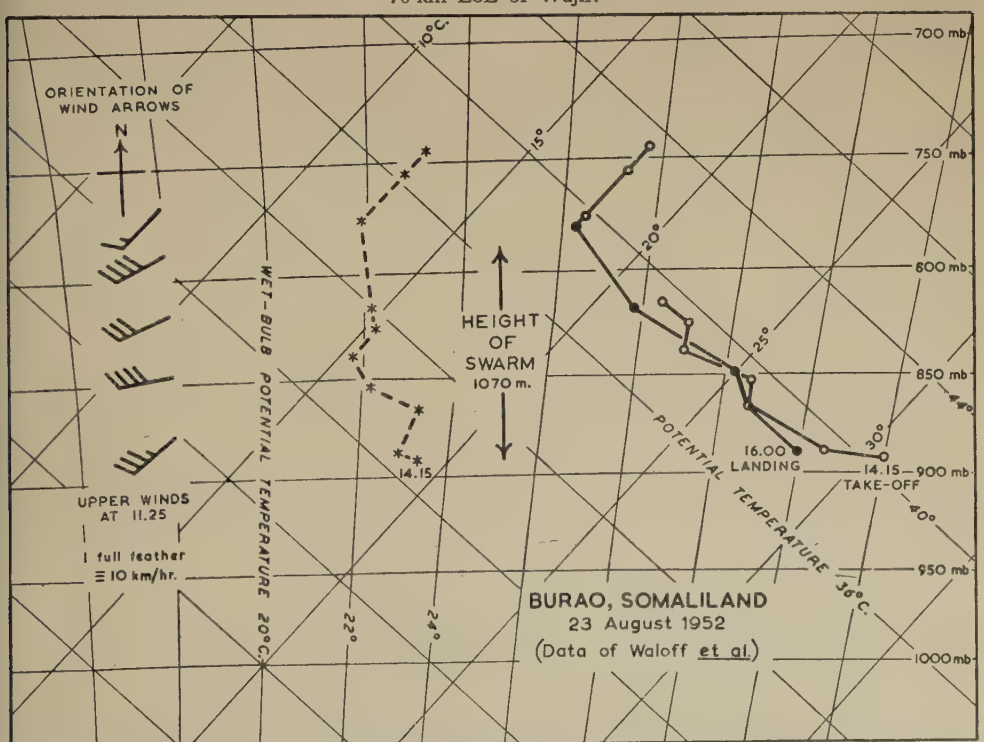

Figure 4. Observations on swarm covering some 20 km², seen between 1430 and 1545 on 23 Aug. 1952, 30 km NW of Burao.

Figures 3 and 4. Upper-air conditions within and around large high-flying swarms. Temperature and humidity recorded by M.O. aircraft psychrometer Mk. III, mounted on starboard jury-struts of Auster observation aircraft, with appropriate corrections for aircraft speed and thermometer lag (Meteorological Office 1945).

at the aircraft at the time, while the corresponding surface reading at Berbera, on th coast 70 km east of Bulhar, was 40°C, and experience in this area indicates that appreciab higher temperatures are likely to have been reached at the same time further inland At 1300 on 26 Nov. 1947 a series of scattered locusts was encountered over northern Uganda by a York aircraft at 3,000 m above sea level, with a true air temperature of 7°C(the corresponding surface temperatures at the time were 29°C both at Kitgum (900 m above sea level) and at Gulu (1,110 m), and 31°C at Lira (1,085 m) and at Soroti (1,110 m Thus both these records of high-flying locusts were associated with lapse rates clos to the dry adiabatic, extending from the ground up to at least the level of the flying locusts which are accordingly likely to have been still under the influence of convection currents from the surface.

Records of stratiform and cumuliform swarm structure during earlier field studies i East Africa, when aircraft observations were not available, had previously been compared with the probable vertical extent of convection for a series of occasions on which this could be estimated from surface observations (Rainey and Waloff 1951). The criterion applied in selecting these cases was the presence of fair-weather cumulus, stratocumulus or thin stratus, all with a vertical extent likely to have been small compared with the corresponding height of cloud base, which was estimated, from the dry- and wet-bulb temperatures recorded at the surface, as the amount of adiabatic ascent necessary to cool the air sampled to the point of condensation (cf., e.g., Fox. 1939). Of 20 such occasions on which observations were made on flying swarms during 1946-47, cumuliform swarm structure was recorded 12 times, with estimated cloud-base levels ranging from 1,000 m to 2,200 m above the ground. The remaining eight occasions, on which swarm structure was recorded as stratiform, gave corresponding estimates of cloud base between 500 and 900 m. It was therefore concluded that convection currents reaching a substantial height suggested as more than 1,000 m, are involved in the development of cumuliform swarm structure.

Subsequent observations have fully supported this conclusion, but have also included repeated records of small swarms, of the order of a square kilometre or less in extent, exhibiting typical stratiform structure despite conditions of vigorous convective turbulence. Thus, for example, a swarmlet some 0·4 × 0·1 km in extent, observed simultaneously from the ground (Z.W.) and the air at Isiolo airfield, Kenya (1205-1245, 6 Feb. 1951), showed extreme stratiform structure, with most of the flying locusts within 10 m of the ground, and only a few as high as 15 m – despite an estimated cloud-base level of 2,400 m above ground (4/8 cumulus, with surface dry-bulb temperature 31°C and wet-bulb 18°C), indicating conditions very similar to those in which cumuliform swarm structure had been observed, for example, in the same area and season just four years previously (2 Feb. 1947 - Rainey and Waloff 1951).

Furthermore, and again in marked contrast with the large swarms (20-110 km^2 previously mentioned (p. 338) in which the topmost locusts were close to the upper limit of adiabatic lapse rates from the ground, it was noted that the topmost locusts in the four successive small swarms which passed through Wajir district in February 1952, and ranged from 1 to 5 km^2 in extent, were far below this upper limit of adiabatic lapse rates. Thus, for example, on the afternoon of 6 Feb. 1952, when flying locusts in a swarm covering 5 km^2 were observed from the air to extend up to 600 m above the ground (the maximum flying height observed in these particular swarms), upper-air temperature observations showed superadiabatic or dry adiabatic lapse rates – and, by implication, convection currents – up to at least 1,500 m from the ground. The same effect was twice observed on a larger swarm traversing the mountainous terrain north of Borama in Somaliland in early August 1957, with an area which appeared to vary from 15 to 35 km^2; at 1210 on the 6th, and at 1320 on the 8th, dry adiabatic lapse rates were found to extend up to more than 400 m above the topmost locusts, which were in turn 150-1,000 m above the ground. On both these occasions direct evidence of vertical air movements above, as well as below the level of the topmost locusts, was secured by accelerometer (p. 346).

Again, small stratiform swarms have been seen on a number of occasions in the vicinity of larger swarms showing typical cumuliform structure, when the larger and smaller swarms are likely to have been experiencing similar conditions. Thus for example at 1635 on 19 Jan. 1954, a small stratiform swarm, 1·2 × 0·4 km in extent, was observed from the air near Kitui, in the Southern Province of Kenya, with many settled locusts, and the topmost flyers only 15 m or so above ground level (*c.* 1,100 m A.M.S.L.); while 30 min later, only 45 km away and over terrain of similar altitude, a larger swarm, 10 × 8 km in extent, was observed flying in cumuliform formation, with the topmost locusts more than 450 m above the ground.

Some direct evidence on the behaviour of flying locusts in a small stratiform swarm under conditions of intense convective turbulence is provided by an observation for which I am indebted to Mr. D. J. McDonald. At about 1700 on 29 Jan. 1956 a swarmlet, of the order of 0·1 km^2 in extent, was seen at Garissa, Kenya, travelling northwards with a light southerly wind, and showing typical stratiform structure, with an even upper surface 6-10 m above the ground, when a small dust-devil arose close to the south-western edge of the swarm. Locusts were rapidly carried up to 40-60 m or so, but quickly descended again, apparently by gliding, into the now-straightened edge of the body of the swarm, as the dust-devil disappeared; the swarm was still flying strongly an hour later.

It would thus appear that some minimum size of swarm, perhaps of the order of 10 km^2 over flat country, may be necessary for the fullest development of cumuliform swarm structure. Furthermore, in addition to the limited heights of flight, relative to the corresponding vertical extent of convection currents, which were thus recorded in these generally smaller swarms, such swarms were also found to exhibit speeds of displacement which were low in relation to the wind speeds concerned and to the corresponding performance of larger swarms (Rainey, in preparation).

Thus the Isiolo swarmlet of 6 Feb. 1951 (p. 340), covering about 0·03 km^2, was travelling directly down-wind at a ground-speed of 3 km/hr in a corresponding wind of 11 km/hr; many locusts were seen temporarily settled beneath the swarm. Again, at Wajir in Feb. 1952, where the track of the small (1-5 km^2) swarms studied was within 1° to 8° of directly down the corresponding vectorial mean wind between the ground and the level of the topmost locusts (on all four occasions on which satisfactory successive swarm fixes were available, together with pilot-balloon observations within less than 50 km of the swarm – Rainey and Sayer 1953), the swarm ground-speeds recorded on these occasions (3 to 8 km/hr) were similarly only 10-30 per cent of the corresponding wind speeds. On the other hand, the higher-flying and much larger swarm (110 km^2) which passed through Wajir district on 12-13 Jan. 1954 (Fig. 3) made good a displacement of 43 km, on a track of 235°, between 1200 and 1520 on the 12th, giving a ground-speed of 13 km/hr., with a corresponding wind of 14 km/hr from 023° (vectorial mean wind between surface and 1,700 m at 1205 at Wajir, some 60 km away). Furthermore, similar track and wind data had been secured for two other large swarms, 150 km^2 and 40 km^2 in extent and both with locusts flying up to more than 1,200 m above the ground, as they passed through the same area during the previous week. The ground speeds found for these swarms were respectively 12 and 14 km/hr, with corresponding wind speeds of 12 and 13 km/hr; the directions of displacement recorded were within 6° and 20° of the corresponding wind directions. Thus, while none of the smaller and lower-flying swarms were travelling at more than 30 per cent of the corresponding wind speeds, all three of the large swarms studied were travelling at ground speeds between 93 and 108 per cent of the corresponding wind speeds recorded. The passage of these large flying swarms was nevertheless still observed (Z.W.) to be accompanied by the temporary settling of large numbers of locusts; and it would therefore appear that it is the higher-flying component, effectively randomly orientated, which acts as a pace-maker for the locusts while they are settling and taking off beneath the swarm.

The swarm observed among the mountains 25-30 km NW of Borama on 6 Aug. 1957, which was relatively large (15-35 km^2) but of relatively limited vertical extent (p. 340), likewise made good a ground speed of only 4½ km/hr between 1200 and 1430,

to the east, with a corresponding westerly wind of 21 km/hr between the surface a 1,000 m at Borama at the time.

Two probably comparable types of travelling swarm have been similarly distinguish by Casimir (1958) in the Australian Plague Locust (*Chortoicetes terminifera* Walke the first type high-flying, rapidly moving (upwards of 150 km/day) and relatively ra and the commoner second type low-flying, generally below 30 m, and slow movin less than 8 km/hr.

These differences in performance between small and large swarms, and observatio such as that on the Garissa swarmlet and the dust-devil (p. 341), suggest the possibili that swarm size (perhaps influencing degree of mutual stimulation) may have some dire effect on the behaviour of their constituent locusts, such as, for example, on the durati of the successive separate flights made by the individual locusts, influencing in turn th scale of the turbulence likely to be encountered in the course of such flights. Anoth effect which may be involved is the contribution made by the heat exchange of the locus themselves to the development of convective turbulence within the swarm, to whic K. F. Sawyer has drawn attention. Thus an examination of field and laboratory evidenc (Rainey, Waloff and Burnett 1957) indicates that a flying locust in sunshine loses hea to the air at a rate of the order of 0·1 cal/sec, of which about half is derived from th respiratory oxidation of fat (Weis-Fogh 1952) and about half from the solar radiatio absorbed by the locust. If most of this heat were retained within the system (as migh perhaps occur in the inner part of a large swarm travelling with the speed of the wind then, at a volume density of 0·1 locust/m^3 (roughly representative of the denser part of the high-flying swarms studied), the temperature of the air within the swarm could b expected to rise by about 0·1° in 3 hr relative to the outside air temperature. Now th effect of a given temperature excess in promoting the vertical displacement of a parce of air increases with the size of the parcel. It has also been pointed out (H.J.S.) that th heat imparted to the air in this manner by a flying swarm must be continuously dissipate across the perimeter of the swarm by atmospheric turbulence (even if the swarm is travelling with the speed of the wind), so that, other things being equal, a larger temperature excess could be expected in the interior of a large swarm than inside a small swarm, with its higher ratio of perimeter to area. Moreover, small swarms with their topmost locusts well below the upper limit of thermal up-currents from the ground, must also lose heat by convection through their upper surface. The importance of the contribution made by the heat exchange of the locusts themselves to the development of convective turbulence could accordingly be expected to increase with the size of the swarm, and might perhaps be appreciable in the case of very large swarms.

All the observations so far considered in relation to vertical temperature distribution and height of flight were made in rainless and mainly fair to fine weather. There is also some evidence of an association between rain and stratiform swarm structure. Thus for example during the 1946-47 observations, apart from the 20 records of swarm structure already mentioned (relating to conditions in which cloud-base could be estimated from surface data), 10 further observations on flying swarms were made in the presence of precipitation, and all recorded stratiform swarm structure (Rainey and Waloff 1951). A reduction in the intensity of convection currents from the surface, after rain, has been suggested on a number of occasions by gliding experience in the Transvaal; conversely, Wills (1943), for example, has suggested an association of particularly vigorous convection currents over England with dry spells such as that of March-April 1938. Consideration of the heat exchange at the ground surface indicates the marked extent to which the energy normally available for convection is likely to be reduced after rain in the arid regions particularly frequented by Desert Locust swarms, where evaporation from a free water surface greatly exceeds the corresponding rainfall. Thus at Garissa, a station representative of much of the invasion area of the Desert Locust in East Africa, a total annual evaporation of more than 250 cm has been recorded with a corresponding rainfall of less than 30 cm (unpublished data of East African Meteorological Department); and Khartoum, which is representative of an extensive area of the central Sudan very

frequently infested by swarms, experiences a mean annual rainfall of 16 cm with a corresponding evaporation of 280 cm from an open water surface (Hurst 1952). In such areas the ground must therefore be dry, for heat-balance considerations, for most of the year, and the proportion of the incoming solar radiation usually available for convection – shown for example by Albrecht (1940) in the eastern Gobi desert to amount to 68 per cent over a year with a rainfall of 24 cm – will accordingly be large. On the other hand, when rain does fall, evaporation from the wet soil surface is likely to account for a substantial proportion of the incoming radiation (45 to nearly 60 per cent in unpublished observations by H. C. Pereira at Kabete in the Kenya highlands); and the energy available for convection must accordingly be drastically curtailed.

4. Direct observations of convective turbulence within and near flying swarms

The first attempt at roughly quantitative observations on the vertical component of air movement in relation to swarm structure was made in Somalia in 1947, by means of special tailed pilot-balloon ascents.

The accuracy of individual balloon fixes by the tail method is limited by the accuracy with which the apparent tail length can be read from the graticule of the theodolite, and by systematic errors due to swinging. Successive observations have accordingly to be smoothed (Meteorological Office 1944), but this smoothing must be done with particular caution when data on vertical air movements are required. As a guide to the extent to which individual tail observations may be in error, tail readings were recorded in the course of a series of twin-theodolite pilot-balloon ascents made under conditions typical of the interior of Somalia (p. 346). It was for example found that at heights of less than 300 m, half the tail observations gave heights within 20 m of those given by the corresponding double-theodolite readings, while at heights of 600-1,200 m the corresponding range of error, covering half the number of tail observations, was 58 m (Rainey and Waloff 1951).

At 0715-0722 on 12 Oct. 1947 a tailed pilot-balloon ascent was made at Garoe, in the Mudugh province of Somalia, within 8 km of a large Desert Locust swarm in which considerable numbers of individual locusts and occasional thin groups were already in intermittent flight up to a height of 5 m or so (Z.W.). A wind of 3 m/sec from 230° was found near the surface, veering and freshening slightly with height to 245° 4 m/sec at 700 m, and the balloon entered 8/8 stratocumulus at 730 m above the surface without having encountered any appreciable vertical air movements; its mean rate of ascent to 690 m (the height corresponding to the last tail observation) was 1·97 m/sec, as compared with the value of 2·00 m/sec to be expected in still air from the weight and free lift of the balloon used.

The flight activity of the locusts increased during the morning, particularly after the cloud began to break at 0915. The height of the topmost flying locusts was estimated at about 15 m at 0855, while from 1130 onwards, until the last of the swarm left the roosting-site at about 1400, locusts were seen overhead from time to time up to the limit of visibility, as specks of light against the blue sky as their wings flashed in the sun, indicating a height probably of the order of 1,000 m. The next pilot-balloon ascent was, unfortunately, not made until 1655, when convective activity is likely to have been decreasing again, but an up-current averaging 4 m/sec was encountered between 200 and 500 m, and one of nearly 2 m/sec between 1,400 and 1,800 m, with no further vertical movement demonstrable above 1,900 m; the corresponding horizontal winds were 5 m/sec, backing gradually from 345° at the surface to 290° at 2,200 m (probably the upper limit of the day's convection), and backing and freshening abruptly with further height to 240-260° 7-8 m/sec at 2,400-2,800 m.

The vertical extent of the swarm was observed to be increasing rapidly while many of the locusts in flight were gliding, but no single gliding locust has yet been clearly observed for long enough (at Garoe or elsewhere) to be certain that it was in fact gaining height – though individual locusts have since been observed gliding, at heights estimated as more than 30 m, for periods of up to 59 sec (1300 19 Dec. 1955; near Lith, Saudi Arabia – a record for which I am indebted to Mr. J. Roffey). It is accordingly of interest

to compare the vertical velocities of the up-currents encountered with what is known of the sinking-speed of the gliding locust in still air. From wind-tunnel measurements Jensen (1956) determined the best lift/drag ratio of a representative Desert Locust with fully-extended wings, as 6·5 at an air speed of 3·8 m/sec, corresponding to a sinking speed of 0·6 m/sec. Gliding locusts in the field, however, have given somewhat higher air speeds (4 to 6 m/sec – Rainey and Waloff 1951), and have been observed (visually and photographically) to have the hind wings partially folded with their leading edges swept back by about 40°, while the forewings were fully extended, in line with each other. A somewhat higher sinking-speed, of about 1 m/sec under field conditions, is accordingly suggested. An up-current of 4 m/sec was in fact recorded at Garoe, at a height well below the probable level of many of the flying locusts; and, while no individual pilot-balloon ascent can adequately sample the range of vertical velocities likely to be experienced at any one time by flying locusts within such a swarm, even the single ascent made on this occasion clearly demonstrated the presence of up-currents sufficiently strong to cause gliding locusts entering them to gain height without muscular effort.

A marked increase in convective turbulence as the cloud began to clear was indicated by the amplitude of short-period fluctuations of air temperature, of which rough estimates were provided by an improvised high-speed thermograph.

A standard Casella thermohygrograph was modified, by changing the gear ratio of the clockwork drive, to give a chart speed of about four times that of the usual daily record. The instrument was housed in a small M.O. pattern screen in which additional ventilation, to reduce lag and render it relatively independent of wind, was provided by a battery-driven fan, installed with its axis vertical, the air intake immediately above the thermograph element, and the outlet discharging beneath the outer roof of the screen, maintaining an air flow of about 3½ m³/min.

From 0535 until 0916, with a continuously overcast sky, the difference between the highest and lowest temperature recorded during each successive period of 14·6 min averaged 0·1°C, with a range of 0·0°C to 0·2°C. Between 0916 and 1211, as 7+/8 Sc lifted and cleared to 1/8 Cu, the amplitude of the corresponding indicated temperature fluctuations averaged 0·8°C, with a range of 0·3°C to 1·4°C. While these figures necessarily underestimate the actual range of air temperature, there can be little doubt of a real change in this respect as the sun broke through, and swarm structure changed from stratiform to cumuliform. Further evidence of increased convective turbulence was provided by a dust-devil seen at 1021.

Fuller evidence on the atmospheric turbulence associated with cumuliform swarm structure has subsequently been secured by the use of a counting accelerometer, installed in light aircraft used for reconnaissance and observation during experimental spraying operations against flying swarms (e.g. Rainey and Sayer 1953).

The instrument, which was developed at the Royal Aircraft Establishment for investigations of fatigue in aircraft structures (Taylor 1952), records the number of times when the upward and downward accelerations, which are imparted to the aircraft in flight by air turbulence, pass thresholds of 0·1 *g*, 0·2 *g*, 0·3 *g*, etc. above and below the 1 *g* datum. The acceleration due to a particular vertical current depends on the vertical velocity of the air current, on the gradient of this velocity which is traversed on entering the current (the 'gust shape'), and on certain characteristics of the aircraft. Current engineering practice considers an arbitrary standard gust-shape in which the vertical velocity increases linearly over a distance of 100 ft to a steady value; and the change in acceleration, in units of *g*, on meeting a vertical gust increasing linearly from 0 to U ft/sec. Equivalent air speed in 100 ft, is given by :

$$\frac{\rho a F U V}{2W}$$

where

ρ is air density, in slugs/ft³

a is $\frac{5{\cdot}73}{1 + 2/A}$, where A is aspect ratio of aircraft

F is gust alleviating factor

V is forward velocity of aircraft, in ft/sec EAS

W is wing-loading, in lbs/ft²

Some indication of the relationship between vertical air velocity and aircraft acceleration, at the lower end of the range of turbulence studied, was provided by a short series of parallel observations with a sensitive rate-of-climb indicator under conditions of minimal turbulence – at 10,000 ft A.S.L., above extensive stratocumulus associated with a temperature inversion of 2°F extending from 7,200 to 8,000 ft, in the vicinity of Nairobi (5,500 ft) at 0745-0845 on 6 July 1951. An initial 28-min normal flight under these conditions gave one gust of + 0·1 *g* and one of − 0·1 *g* only, and during a further 16-min run, with the aircraft trimmed with particular care for level flight, the accelerometer recorded no gusts at all. This suggests that the degree of turbulence encountered may not have been far below the threshold of sensitivity of the instru-instrument, for which the expression quoted gives a figure of 2·8 ft/sec as equivalent to 0·1 *g* under these conditions. During this 16-min period some 74 instantaneous readings of the rate-of-climb indicator were made at 10-sec intervals. Of these, 38 were recorded as 0, 65 were between + 1 and − 1 ft/sec inclusive, 73 were between + 2 and − 2 ft/sec, and there was a single reading of 3 ft/sec, which is reasonably consistent with the computed threshold of sensitivity of the accelerometer under these conditions.

Fig. 5 shows the gust-spectra so recorded among high-flying locusts on two different occasions. The first, in Kenya in 1952, relates to a swarm covering $3\frac{1}{2} \times 2$ km, with locusts flying up to about 600 m above the flat and arid thorn bush of the Northern Frontier Province. Three successive observation flights, accompanying three spray-sorties, gave similar degrees of turbulence, and their records, taken mainly between 300 and 600 m above the ground, have accordingly been combined. Of this total accelerometer record of 119 min, more than 70 min were spent within a kilometre or so of the swarm, including repeated passages through it among the flying locusts; the remainder of the period, less than 50 min, covers the time spent in flight between the swarm and the base airstrip at Wajir. Unfortunately up-gusts exceeding 1·3 *g* could not be separately recorded, owing to accidental damage to the instrument. The second occasion, in Somaliland in 1957, relates to a swarm (already mentioned, p. 341) covering $8 \times 5\frac{1}{2}$ km, with locusts flying up to 1,000 m above the ground, over the broken western slopes of the Simudi mountains in Borama district. Accelerometer records were secured within the swarm in the course of two traverses, at heights varying from about 100 to 500 m above the ground, over the total period of 200 sec during which the aircraft was in physical contact with flying locusts. A very similar degree of turbulence had been recorded in the vicinity of this swarm, below the level of the topmost locusts, on each of the two preceding days (1039-1043 and 1226-1235 6 Aug.; 1219-1249 7 Aug. 1957).

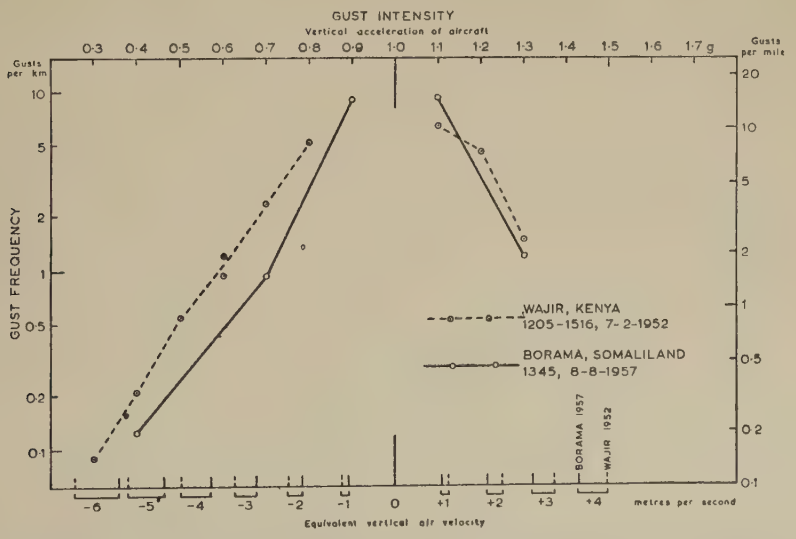
GUST INTENSITY
Vertical acceleration of aircraft
0·3 0·4 0·5 0·6 0·7 0·8 0·9 1·0 1·1 1·2 1·3 1·4 1·5 1·6 1·7 g
Gusts per km
Gusts per mile
GUST FREQUENCY
10 5 2 1 0·5 0·2 0·1
20 10 5 2 1 0·5 0·2 0·1
WAJIR, KENYA
1205-1516, 7-2-1952
BORAMA, SOMALILAND
1345, 8-8-1957
BORAMA 1957
WAJIR 1952
-6 -5 -4 -3 -2 -1 0 +1 +2 +3 +4
metres per second
Equivalent vertical air velocity

Figure 5. Atmospheric turbulence among flying locusts in cumuliform swarms. Gust-spectra recorded by counting accelerometer in Auster JI and JIb aircraft; owing to differences in all-up-weight and, in turn, of wing-loading (9·2 lb/ft² at Wajir and 10·8 at Borama), the relationship between vertical acceleration and equivalent air velocity was different on the two occasions.

Fig. 5 indicates that, in an average kilometre on both occasions, up-currents exceedin 1 m/sec were encountered at about seven points, of which about three exceeded 2 m/sec The accelerometer data, if they may be accepted as a representative and effectively instan taneous record of the degree of turbulence within these swarms, thus indicate that gusts equivalent to up-currents of 1 m/sec (the approximate sinking speed of gliding locusts p. 344) or more, are likely to have occurred among the higher flying locusts at some 5 points per km^2 at any one time, and, of these, about 10 could be expected to exceed 2 m/se

The accelerometer data on turbulence within these swarms thus provide evidence o up-currents of a roughly appropriate order, both in strength and spacing, to account fo effects such as the pillars of locusts recorded for example on 28 Sept. 1946 (p. 337). I would however appear that such locust towers may – like small cumulus clouds – outliv the particular up-currents which produce them, since observations of a sensitive rate-of climb indicator, during traverses of billowing projections from the upper part of a larg cumuliform swarm (*c.* 1200 12 Jan. 1954 – Fig. 3), provided no evidence of any marke associated up-current. Moreover, during the Borama observations, marked turbulence with adiabatic lapse rates (p. 340), was found to extend well above the level of the topmos locusts. Thus records taken at 0-400 m above this swarm (i.e., at 2,000-2,400 m A.S.L.), a 1204-1213 on 6 Aug. and at 1317-1323 on 8 Aug., gave frequencies of the lighter gust (those counted only at 0·9 and 1·1 *g*) which were still 80-81 per cent of the correspondin values recorded below the level of the topmost locusts immediately afterwards; and stronger gusts, counted at 0·7 and 1·3 *g*, were also still encountered above the top of the swarm, though only about a third as frequently as below this level.

Air turbulence may also directly limit the efficiency of application of insecticide, and particularly the use of fine sprays against tropical ground targets such as settled locusts. A criterion of intensity of turbulence is therefore necessary for the adequate specification of meteorological factors affecting such operations (Yeo 1954). Examples of the utilization of counting accelerometer records for this purpose are recorded elsewhere (Rainey 1958).

Further evidence of the strength and distribution of convection currents likely to be encountered by flying swarms was provided by a series of twin-theodolite pilot-balloon ascents (to which brief reference has previously been made – Rainey 1948b), carried out for this purpose in Somalia during January 1948, at Belet Uen, which is representative of an extensive area of flat and almost featureless terrain which is subject to particularly high swarm incidence at this season of the year. For obtaining data on vertical air movements at relatively low levels, low-lift balloons were used (standard 70-in. balloons, of mean weight 21 g, inflated to a free lift of 5 *g*). 65 such ascents were completed, with 10-16 ascents during each of the five 2-hr periods between 0600 and 1600; I am greatly indebted to the Director-General of the Meteorological Office for providing assistance in the computation of the results. The balloon heights, as recorded 30 sec after launching, were examined in the first instance for evidence of a diurnal variation in the variance of these heights, as determined for each successive 2-hr period. Bartlett's test (see, e.g., Brownlee 1949) demonstrated that the variance of these balloon heights recorded between 0600 and 0800 was significantly (P between 2 and 5 per cent) less than that found later in the day, with no significant difference between the four other 2-hr periods in this respect. The observations over the whole period 0800-1600 may therefore be considered together, as representative of higher-turbulence conditions, in contrast with the lower-turbulence conditions illustrated by the 0600-0800 data (Fig. 6.)

The heights recorded at 30 sec during the 49 ascents made between 0800 and 1600 gave a standard deviation of 13·1 m ($\equiv$ 0·4 m/sec), while the corresponding standard deviation found in the 0600-0800, low-turbulence observations was 6·1 m ($\equiv$ 0·2 m/sec), with an overall mean of 32·3 m and standard error 1·5 m. This overall mean is in close agreement with the theoretical value of 31·7 m given by the usual formula $q\,L^{\frac{1}{2}}/(W+L)^{\frac{1}{2}}$ for the weight and free lift used; and the heights recorded in Fig. 6 have accordingly been expressed also as vertical air velocities, on the assumption of a constant rate of ascent of 1·06 m/sec relative to the air being traversed by the balloon.

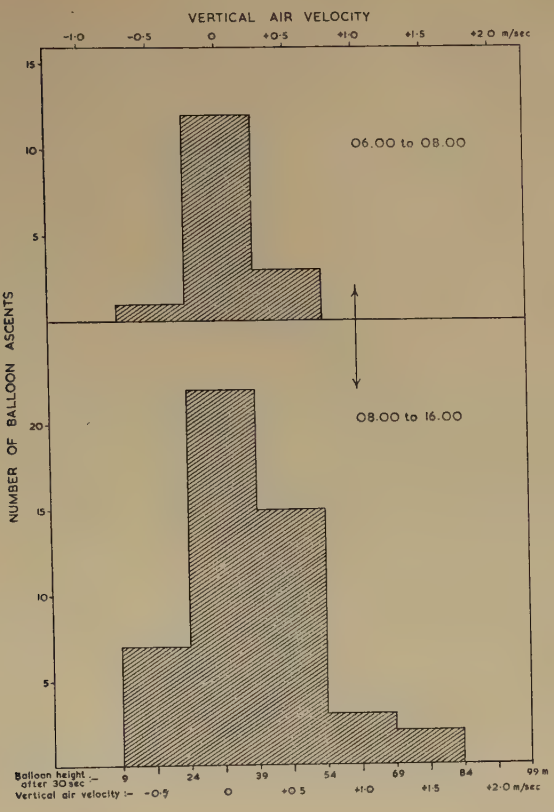
VERTICAL AIR VELOCITY
-1·0 -0·5 0 +0·5 +1·0 +1·5 +2·0 m/sec
NUMBER OF BALLOON ASCENTS
06.00 to 08.00
08.00 to 16.00
Balloon height after 30 sec :-
9 24 39 54 69 84 99 m
Vertical air velocity :- -0·5 0 +0·5 +1·0 +1·5 +2·0 m/sec

Figure 6. Vertical components of air movement at low levels over central Somalia. Frequency distribution of heights attained by low-lift pilot balloons 30 sec after release, expressed as vertical air velocities corresponding to a rate of ascent of the balloon of 1·06 m/sec in still air (see p. 346); Belet Uen, 9-18 Jan. 1948. Up-currents exceeding the probable sinking-speed of gliding locusts (1 m/sec, indicated by double-headed arrow) were encountered, at a height of about 20 m, during 4 out of 49 such ascents made between 0800 and 1600, but were not recorded during 16 such ascents made between 0600 and 0800.

Since each of these figures for vertical air velocity is a mean value over the vertical distance between ground level, at which vertical air velocities must be zero, and the height reached at 30 sec, this mean value must necessarily underestimate the vertical air velocity experienced by the balloon at 30 sec, but may be regarded as an approximate estimate of that experienced at 15 sec. Since the timing of the balloon launches may be regarded as random with respect to the airflow, the figures constitute a sample which could be expected to be reasonably representative (see below) of the vertical air velocities existing over the area. Fig. 6 thus suggests that under such conditions nearly 10 per cent of uniformly-distributed locusts in flight at heights around 20 m at any one time are likely to be experiencing up-currents exceeding their gliding sinking-speed.

At greater heights, larger vertical air velocities were recorded : up-currents between 4 and 6 m/sec were for example encountered during six of these ascents, at heights of 300-800 m. No unbiased direct estimate of the relative frequency of occurrence of such up-currents is however available, since as soon as a pilot-balloon is released its sampling of vertical air velocities must begin to be biased by the associated convergent or divergent nature of the horizontal flow pattern through which it rises. This effect must be expected to cause an initial overestimation of the frequency of up-currents, to an extent likely to increase initially with height, and to be more marked with low-lift than with high-lift balloons.

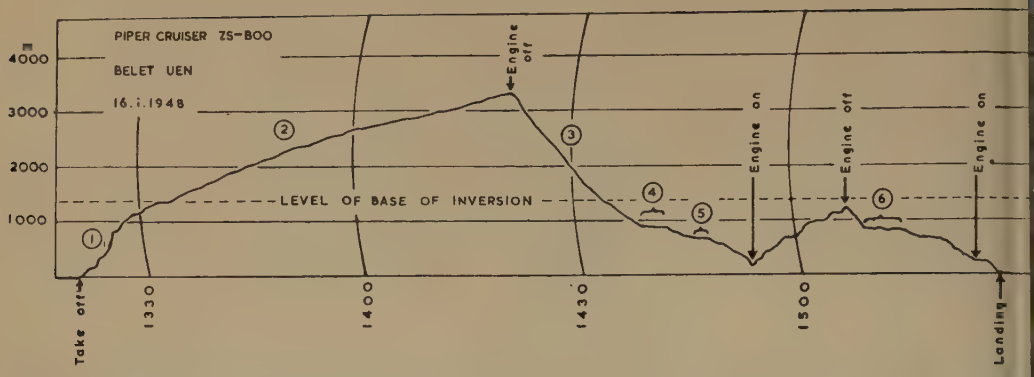

Figure 7. Evidence on thermal up-currents provided by light aircraft. Barograph record of flight showing (1) initially increased rate of climb, by utilization of thermals below a marked inversion (Fig. 8), then (2) steady rate of climb above inversion, to 3,350 m, followed by engine-off descent in which an attempt was made to utilise all thermals encountered. Above the inversion, no up-currents were detected (3); below the inversion it was found possible to maintain height, by circling in up-currents, during successive periods of 4 min (4) and 2 min (5). The engine was then used to climb back to 1,200 m, and was then stopped again; during the following descent, a further thermal was encountered, in which height was maintained for 4·9 min (6).

Neither pilot-balloons nor accelerometer provide any direct information on the horizontal extent of individual up-currents. Some evidence on this point (Rainey 1948a) was secured in the course of one of the aeroplane ascents undertaken to provide upper-air temperature data during the Belet Uen observations. On this occasion (Fig. 7) it was found possible to maintain an altitude between 760 and 790 m above the ground for 4·9 min, with the engine switched off, by circling continuously. The sinking-speed in still air of the aircraft flown in this manner had been determined directly earlier in the same flight (above a well-marked temperature inversion at 1400 m – Fig. 8) as 3·7 m/sec; and the diameter of the circles flown was 0·4 km. A continuous up-current, averaging more than $3\frac{1}{2}$ m/sec, must accordingly have extended over an area at least 0·4 km in diameter for a period of nearly 5 min. There is, however, not much evidence as to the relative frequency of occurrence of up-currents of such extent. During the 36 min of this particular flight made without engine and below the inversion, height was similarly maintained on two other occasions for shorter periods (4 and 2 min), making a total of 11 min out of 36; but sampling was highly selective, in that the strongest up-currents were actively sought, by means of the sensitive altimeter, and were, moreover, encountered above the settlement of Belet Uen itself, where the distribution of vertical air movements may well have differed from that over the surrounding bush.

Such up-currents are actively exploited by gliding birds, as a means of sustained flight without muscular work; and the particular up-current just mentioned was shared with a number of large soaring birds (vultures and Marabou storks), circling in company with the aircraft. It is not certain whether locusts may similarly make active use of these vertical air movements, or are merely passively displaced by them. The facts that the topmost flying locusts are often well below the upper limit of thermal up-currents from the ground (p. 340), and that, even when locusts have been seen to attain this limiting level, they have done so only over an area instantaneously very small compared with the area of the base of the swarm (p. 338), somewhat discount the possibility of deliberate utilization of up-currents, and suggest that the topmost locusts may usually be descending relative to the air (gliding ?) On several occasions (e.g., Mtito Andei, Kenya; 1520/10 and 1715/11 Feb. 1955), groups of the order of a hundred storks, accompanying swarms, have been seen with the lowest birds, in soaring flight, at roughly the level of the topmost locusts, 700-900 m above the ground. Such differences in height of flight would be consistent with a deliberate utilization of up-currents by the birds but not by the locusts. A neuro-physiological test of a suggested sensory mechanism for the direct perception of up-currents by the locust has given a negative result (Pumphrey, *in* Rainey and Waloff 1951).

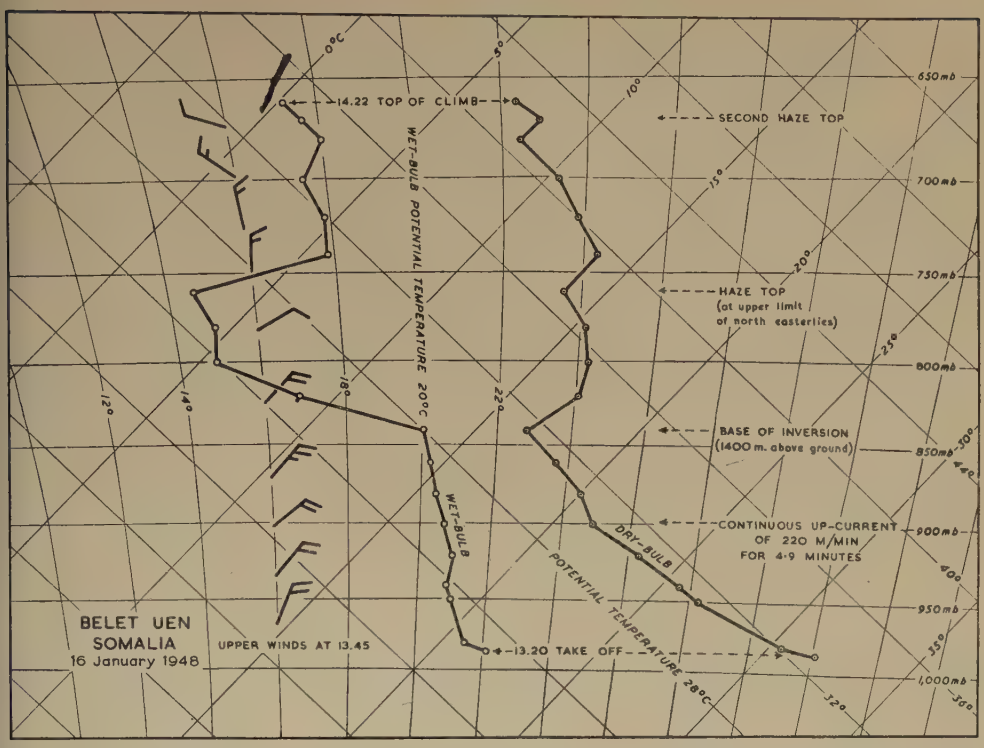

Figure 8. Upper-air conditions during flight recorded in Fig. 7. Temperature and humidity by aircraft psychrometer Mk. III, corrected for airspeed and lag as in Figs. 3 and 4; upper winds by twin-theodolite pilot-balloon observations; NE monsoon (cf. Fig. 3).

W. G. Penney has directed attention to the extent to which the disturbance of the air by the actual flight of the locusts themselves must also contribute to the turbulence developed within a swarm. Thus a swarm with an average area density of 40 flying locusts/m^2 (recorded for example at Kanga in Kenya at 1615-1655 on 10 Feb. 1955), with each locust weighing about 2 g, would necessarily subject the ground being traversed to an additional dynamic pressure of some 0·008 g/cm^2, corresponding, since $\delta p = \frac{1}{2}\rho v^2$, to a down-current averaging 1·1 m/sec over the area of the swarm. The effects of turbulent mixing could be expected to spread the effects of such down-currents somewhat beyond the perimeter of the swarm, over a wider area with a corresponding reduction of vertical velocity, particularly in the case of smaller swarms with their higher ratio of perimeter to area. No particularly striking down-currents have so far been noted during aircraft observations within swarms, but such observations (e.g., Fig. 5) have been mainly confined to the upper parts of swarms, while the down-currents under consideration would be expected to reach their maximum intensity at the lower levels. Observations from the ground, moreover, have several times given an impression of 'avalanches of locusts,' which might be evidence of such down-currents. Thus, for example at Kanga, on the occasion just quoted, the observation point was engulfed by the rapid descent of a curtain of flying locusts so dense as temporarily to obscure completely, within a few seconds, bushes and trees at a distance of less than 30 m (Z.W. and H.J.S.).

Discussion

The foregoing evidence, of marked effects of convective air movements on the structure exhibited by flying swarms of locusts, suggests the converse possibility that observations on flying locusts, like those on herring-gulls (Woodcock 1940a, b), and on other soaring birds and dragonflies (Scorer 1954), might throw some light on details

of the corresponding air movements. In considering such possible utilization of locusts as indicators of airflow, it is of course necessary to begin by realizing that the flying locust is not an inanimate object, and to attempt to assess and to allow for the contributions made by the active behaviour of the locusts themselves to the effects observed. While our understanding of the relevant aspects of locust behaviour is still inadequate for detailed individual assessments of this kind, certain general conclusions are beginning to emerge. Thus while groups of flying locusts, exhibiting a relatively uniform orientation within each group, are commonly seen in flying swarms, and locusts appear in general to take off and to land into wind (Kennedy 1951; Waloff and Rainey 1951), the widest possible diversity of orientation has been recorded between different groups, or from time to time at a single point, within the same swarm (Rainey and Sayer 1953; Waloff 1958); and the data so far available on the direction and speed of movement of individual swarms (p. 341) are all consistent with an effectively random orientation of the flying locusts concerned. Furthermore, there is evidence that much of the close and apparently purposeful association between long-range movements of Desert Locust swarms, and the distribution of the rainfall which is essential for successful breeding, represents in fact downwind displacement (which would necessarily result from random orientation), towards and with zones of convergent surface windflow (Rainey 1951, 1952, 1955).

Perhaps the most striking effect of the active behaviour of flying locusts, on the other hand, operating against the effects of the environment, is the continued cohesion of individual swarms, observed over distances of hundreds of kilometres and times of many days, without appreciable stragglers or dispersion, despite the disruptive action of vigorous convective turbulence – as well as of the apparently random orientation of the flying locusts themselves in the interior of such swarms. Thus for example the extent of one particular swarm, with which contact was maintained for more than a week as it crossed southern Kenya during January 1954, was determined by detailed aircraft traverses on three occasions during this period. At 1200 on the 16th, over the Tana lowlands (*c* 300 m above sea-level) near Garissa, the swarm was observed (H.J.S.) to cover 16 × 5 km, with flying locusts up to 900 m above the ground; three days later, at 1705 on the 19th, over the northern Yatta plateau at about 1,100 m above sea level – after flying for about 30 hr and covering a distance of 155 km – the area of the swarm, observed by the writer, was almost unchanged (10 × 8 km), with flying locusts up to more than 450 m above the ground; and at 1500 on the 25th, south-east of Narok at an altitude of about 1500 m in the Rift Valley – 370 km and about 70 hours' flying from the site of the first assessment, and 9 days later – the swarm was 18 km long and roughly 5 km wide, with flying locusts up to more than 900 m above the ground. The difference of roughly 10 per cent between the recorded initial and final extent of this swarm is likely to have been well within the limits of reproducibility of such traverses. Spraying operations on a limited scale were carried out against the swarm between the 20th and 25th, but, from assessments under comparable conditions, can be expected to have killed only a few per cent of the locusts in the swarm (Rainey 1956, 1958).

The scale of the potentially disruptive effects, both of the diversity of orientation of the locusts, and of atmospheric turbulence, which are successfully withstood by such a swarm, are indicated by the rates of dispersion to be expected, in the first case, of randomly-orientated locusts flying independently of each other, and, in the second case, of a cloud of inert airborne particles, of the same size as the swarm and under similar conditions of turbulence. I am greatly indebted to Mr. E. P. Hicks for the following theoretical treatment of these two cases, and to Mr. J. G. Skellam for the two-dimension value of the ' diffusion ' coefficient used in the first case.

For the first case the ' swarm ' is considered as comprising a large number of locusts (or a large number of groups of locusts), each locust (or group) flying horizontally and independently of the others at constant speed v, in a straight line for an exponentially-distributed average distance L (the ' mean free path '), and then changing direction abruptly to a new course totally uncorrelated with its previous motion. This is, as first pointed out by H. J. Sayer, a ' random walk ' analogous to that considered in the kinetic

theory of gases. If such ' locusts ' are envisaged as having been initially released at a point then, at time t after release, they would be distributed in a circular Gaussian probability distribution centred at the release point. The standard deviation $\sigma(t)$ of the distribution would be given by :

$$\sigma^2(t) = Lvt$$

Such a ' swarm ' would have no sharply-defined boundary; but if the boundary is taken to be at $2\frac{1}{2}$ times the standard deviation, which would include 95 per cent of the ' locusts,' then the diameter d of the ' swarm ' at time t would be :

$$d = 5\,\sigma(t) = 5\,(Lvt)^{\frac{1}{2}}$$

Now such a ' swarm,' of initial diameter d_0, can be imagined as having been formed from a point source at an earlier time t_0, where :

$$d_0 = 5\,(Lvt_0)^{\frac{1}{2}}$$

Then, if time is measured forward from the instant when the diameter is d_0, the diameter d at time t would be :

$$d = 5\,[Lv\,(t + t_0)]^{\frac{1}{2}}$$

$$= d_0\,(1 + 25\,Lvt/d_0^2)^{\frac{1}{2}}$$

For the actual swarm under consideration, d_0 may be taken as 8 km, and field and laboratory determinations of the air speed of flying locusts (Weis-Fogh 1956; Rainey, Waloff and Burnett 1957) indicate that v is likely to have been about 15 km/hr. Considering L, visual impressions, with some photographic data, on the individual groups of flying locusts which have been observed in the interior of such swarms, indicate that a figure of 0·1 km may be regarded as a probably conservative estimate of the ' mean free path ' of such groups. Substituting these values :

$$d = 8\,[1 + 25 \times 0{\cdot}1 \times 15\,t/8^2]^{\frac{1}{2}}$$

$$= 8\,[1 + 0{\cdot}59\,t]^{\frac{1}{2}}$$

This means that a hypothetical ' swarm ' initially 8 km in diameter, and consisting of locusts (or groups of locusts) flying horizontally and independently of each other at 15 km/hr, with each locust (or group) changing orientation at random every 100 m, would expand (in still air) to twice its original diameter within 6 hr; to more than 4 times its original diameter after 30 hours' flight (when the extent of the real swarm was in fact almost identical with its original value); and to $6\frac{1}{2}$ times its original diameter after 70 hr, when the real swarm showed only a doubtful 10 per cent increase in its original extent. Furthermore, even a real increase in diameter by as much as 20 per cent after 70 hr would correspond, in terms of random walk dispersion, to a ' mean free path ' of only 1 m. It is thus clear that mere changes of orientation, even at intervals much shorter than any yet observed in the field, could not account – even in the absence of turbulence – for the degree of cohesion recorded, if such changes of course were made completely at random.

In the second case – the dispersion by atmospheric turbulence of inert airborne particles, which may again be envisaged as having originated at an instantaneous point source – the distribution in space at subsequent times may again be taken as Gaussian (though not spherical). Considering only the cross-wind horizontal dispersion, the cross-wind standard deviation $\sigma(t)$ is given by the empirical formula of Sutton (1953) :

$$\sigma^2(t) = \tfrac{1}{2}\,C^2\,(wt)^{\frac{7}{4}}$$

where C is the appropriate lateral diffusion coefficient and w the corresponding mean wind speed. If, as before, the diameter of the ' swarm ' is taken as 5 σ, and time is measured

from the instant when the diameter is d_0, then :

$$d_0 = (5/2^{\frac{1}{2}})\, C\, (wt_0)^{\frac{7}{8}}$$

and

$$d = (5/2^{\frac{1}{2}})\, C\, [w\,(t + t_0)]^{\frac{7}{8}}$$

so that

$$d = d_0\, [1 + (5\, C/d_0)^{\frac{8}{7}}\, 2^{-\frac{4}{7}}\, wt]^{\frac{7}{8}}$$

C varies with stability, altitude, terrain and possibly other factors. From a recent review of the data available (United States Weather Bureau 1955), a value of the order of 0·11 may be suggested for the conditions of large lapse, heights of up to 1,000 m, and rough terrain which were involved. w was about 5 m/sec, and, taking d_0 as 8,000 m, as before :

$$d = 8{,}000\, [1 + (0{\cdot}5/8{,}000)^{\frac{8}{7}}\, 2^{-\frac{4}{7}}\, 5\, t]^{\frac{7}{8}}$$

$$= 8{,}000\, [1 + 0{\cdot}528 \times 10^{-4}\, t]^{\frac{7}{8}} \text{ metres with } t \text{ in seconds}$$

$$= 8\, [1 + 0{\cdot}19\, t]^{\frac{7}{8}} \text{ km with } t \text{ in hours}$$

This means that the degree of turbulence likely to have been encountered by the actual swarm under consideration could have been expected to disperse a cloud of completely inert airborne particles, in area and vertical extent similar to the swarm, at a rate which would double the cross-wind diameter after about six hours, would increase it to five times its original value after 30 hours' flight, and by ten times after 70 hr. Furthermore, while there is admittedly considerable uncertainty in estimating the exact value of C appropriate to particular conditions, even a real increase in diameter by as much as 20 per cent after 70 hr, by atmospheric diffusion, would imply a lateral diffusion coefficient of only 0·003 – smaller, by more than an order of magnitude, than such values as have actually been determined under roughly comparable conditions (United States Weather Bureau 1955). Field data on this point are illustrated by recent observations in southern England on the distribution of airborne particles from a ground-level point-source, which gave a cloud which spread out to a width of 5-6 km at a distance of 10 km down-wind from the source, and to a width of 27 km at a distance of 64 km (Pasquill 1955).

It is therefore concluded that in such a swarm some behaviour process must operate to oppose the disruptive effects both of atmospheric turbulence and of the diversity of orientation of the flying locusts within the swarm; and that while the data available on the direction of displacement of individual swarms (and, in the case of the larger swarms, on their corresponding ground speeds) are consistent with an effectively random orientation of the flying locusts, such observations of continued cohesion are not. The relatively well-defined edge of such swarms, without indication of any further systematic increase in the density of the locusts towards the centre of the swarm, and the diversity of orientation observed in the interior, all direct attention to behaviour at the perimeter. Following the 1952 aircraft observations, it was realized that the observed continued coherent existence of swarms behaving in this manner would appear to imply the existence of some reaction having the effect of deflecting inwards groups of flying locusts when they reach the perimeter of the swarm. Direct field evidence of such behaviour was subsequently secured (Rainey, Waloff and Sayer 1955; Waloff 1958).

The flying speed of locusts is such that the potential rates of dispersion quoted could be opposed by a relatively small proportion of inwardly-directed flight. If such flight takes place all round the perimeter – as direct observations suggest – it could be without effect on the direction of displacement; and the randomly-orientated element of the flying locusts, in the interior of the swarm, could still determine the direction, and, in larger swarms, the speed, of displacement of the swarm as a whole.

Thus a flying swarm appears to represent a state of dynamic equilibrium between the disruptive effects of atmospheric turbulence and the cohesive behaviour of the flying locusts – reminiscent, to G. I. Taylor, of the opposing effects of gravitation and heat

motion in interstellar gas clouds. The actual behaviour concerned may be regarded as reaction(s) opposing further separation of the locusts reaching the perimeter of the swarm, somewhat analogous to the elementary molecular theory of surface tension*; and it seems possible that similar behaviour may be involved in maintaining locusts, carried aloft by convective up-currents, in the sheet formations which appear to be involved in a number of the characteristic manifestations of cumuliform swarm structure.

There is now some evidence on the sensitivity of the sense organs which are available to provide the instrumentation required for such reactions of flying locusts to each other. Thus neuro-physiological studies of visual acuity in the Tropical Migratory Locust, *Locusta migratoria migratorioides* L., have demonstrated electrical responses in the nerves from the eyes, evoked by movements of a small light source through as little as 0·1° – the angle subtended by the length of a locust at a distance of about 30 m (Burtt and Catton 1956). Furthermore, Pumphrey (1940) found that the sound emitted by a flying locust evoked a demonstrable electrical response from the tympanal organ of a second locust, though subsequent neurophysiological work and field recordings (Haskell 1957) indicate that the distance over which one locust can hear another in flight under field conditions is likely to be only of the order of 2 m or less.

Such, briefly, is the present state of knowledge of the part played, in the structure and movement of flying swarms, by the behaviour of the locusts themselves, which must be borne in mind in any attempt to utilize the locusts as indicators of air movements. Superficially, up-currents in the form of the bubbles postulated by Scorer and Ludlam (1953) would appear to be particularly appropriate for accounting for the foam-like structure suggested by the reticulated effects often observed; while the association, recorded on several occasions, between gusts and temporary apparent increases in the number of locusts seen flying low and settling, may be interpreted as an ascent of locusts during lulls, followed by descent with gust fronts, corresponding with the distribution of vertical air movements within the convection cells postulated by Durst (1932); and eddies on a smaller scale probably account for the rapid swirling movements seen on looking upwards through high flying swarms (p. 337). Objective data on the course, track and ground speed of individual flying locusts in swarms, secured by a technique of double-exposure vertical photography due to Sayer (1956) and now in course of analysis, may provide evidence on some of the still finer structure of turbulent wind flow, though the complete interpretation of these and other observations on flying swarms must await more detailed knowledge of locust behaviour as well as of atmospheric turbulence.

Acknowledgments

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* An analogy suggested by W. Schaffer, and subsequently developed by H. J. Sayer as a convenient method of considering the effects of such individual reactions on the behaviour exhibited by the swarm as a whole.

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The meteorology of North Greenland during the midwinter period

By R. A. HAMILTON
Meteorological Office, Prestwick

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Summary

During the midwinter months the general airflow over N. Greenland is southwesterly, but near the surface of the ice sheet, which is cooled by radiation in the polar night, katabatic winds flow from the higher parts of the ice sheet towards the coast. At sheltered places on or near the coast pools of cold stagnant air collect in fine weather when the pressure gradient is weak.

Analysis of the 700 mb charts for two midwinter seasons shows that breaks in the predominantly fine weather were caused by depressions and their associated fronts which crossed the inland ice. There were about 15 disturbances per month which affected N. Greenland, but only a few of these caused precipitation. Most of the fronts were weak and the snowfall slight. During the first period of two midwinter months under consideration the only considerable fall of snow occurred when a deep depression remained almost stationary over S. Greenland for three days and drew moist air from the N. Atlantic over N. Greenland. The onset of the strong northwesterly wind on the coast after a period of calm is generally associated with the passage of a cold front.

There is normally an inversion in the lowest layers above the ice sheet and the air temperature on the inland ice varies with the wind speed and the air mass. It is estimated that the temperature difference between the bottom and top of the inversion varies from around 0°C in overcast conditions with a strong wind to around 25°C in clear and calm conditions. Warm air is sometimes carried in the southwesterly circulation over the inland ice : the highest temperature measured on the ice sheet in the midwinter months was − 18°C in a warm sector, and the lowest was − 66°C in fine and almost calm weather. On the coast the onset of the northwesterly wind is marked by a sudden rise of temperature and fall of humidity : when the wind falls to calm the temperature falls rapidly at first, but within 24 hours reaches a value from which it differs little during the period of several days before the next onset of the northwesterly wind.

There is a fairly close relationship between the height and intensity of drift snow, the wind speed and the visibility. Snow does not begin to drift until the wind speed reaches 10 kt; the snow drift is moderate with a visibility of about 2 km when the wind speed is 17 kt and is heavy with a visibility of about 200 m when the wind speed is 27 kt. After a spell of fine weather over N. Greenland drift snow was absent on the inland ice even with a wind speed of 21 kt. Drift snow was observed in the coastal area only after a fall of snow.

A comparison of the mean pressures and precipitation totals for the midwinter months under review with mean charts and mean monthly precipitation totals for other stations in the area for different years suggests that the two Januarys studied were normal months, but the two Decembers were abnormal.

1. Introduction

In a previous paper (Hamilton 1958), the meteorology of N. Greenland in the midsummer period was discussed. In this paper the meteorology of N. Greenland in midwinter is studied in relation to the synoptic situation as analysed on the 700 mb charts.

2. Synoptic types

A detailed study has been made of the weather experienced at " Northice "* (78° 04′ N, 38° 29′W, altitude 2,343 m) on the inland ice and at Britannia Sø (77° 09′ N, 23° 26′ W, altitude 229 m) in Dronning Louise Land during the period 1 December 1952 to 31 January 1953. This paper refers mainly to this period, but when this study was complete, a similar, but less detailed, study was made of the period 1 December 1953 to 31 January 1954 to enable a comparison of the two seasons to be made.

The Climatological Tables (Hamilton and Rollitt 1957) show that at midwinter at Northice there was normally a westerly wind of about 16 kt and a sky which was usually almost clear but sometimes overcast, with a mean cloud amount of less than 3 oktas. At Britannia Sø the sky was normally cloudless, but on occasions there was cloud sometimes with snowfall. There was normally either a strong NWly wind from the inland ice, or else the wind was almost calm. The changes from the normal fine weather and the onset of NWly wind from the inland ice were usually indications of frontal activity.

* " Northice " is not an accepted place name, but is used in this paper for convenience.

The general direction of airflow at 700 mb was SWly but other types persisted fo some days. During the two-month period the types of airflow were as follows :

1 to 8 Dec.	SWly airflow
9 to 13 Dec.	NWly airflow followed by ridge
14 to 15 Dec.	Wly airflow
15 to 17 Dec.	Ridge of high pressure : pressure gradients small
18 to 25 Dec.	Sly airflow
26 Dec. to 7 Jan.	SWly airflow
8 to 9 Jan.	NWly airflow
10 to 13 Jan.	SWly airflow
14 to 17 Jan.	Variable
18 to 20 Jan.	Ridge of high pressure : pressure gradients small
21 to 24 Jan.	SWly becoming Sly airflow
25 to 28 Jan.	SE to Ely airflow
29 to 31 Jan.	Ridge of high pressure : pressure gradients small

The mean 700 mb charts for December 1952 and December 1953 (Figs. 1 and 2 show low pressure over the Canadian Archipelago and high pressure over the N. Atlantic with a general southwesterly airflow over N. Greenland, though it is more westerly in December 1952 and more southerly in December 1953. Both months will be shown in § 8 to have been rather exceptional.

3. Determination of altitude of 700 mb surface at Northice

The measured pressure at Northice was always greater than, but not much greater than 700 mb, so the station was always below, but not much below the 700 mb constant pressure surface. The height above m.s.l. of the 700 mb surface at Northice can therefore be determined from the pressure readings and an estimate of the mean temperature of the air column between the barometer and the 700 mb surface.

No measurements were made during the winter at Northice of the temperature of the air above screen level, so the difficulties of the estimation of the mean temperature of the air column between the barometer and the 700 mb surface were dealt with as follows.

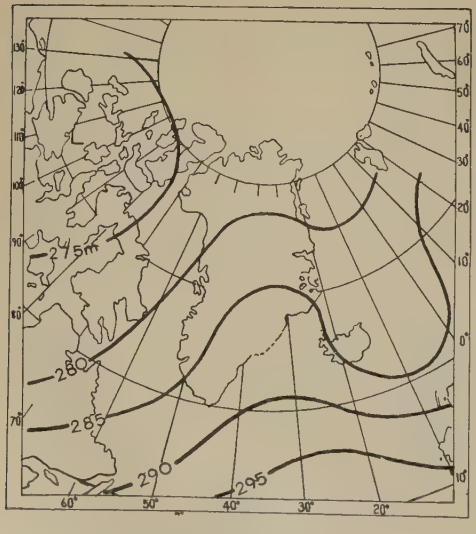

Figure 1. Mean Dec. 1952, 700 mb.

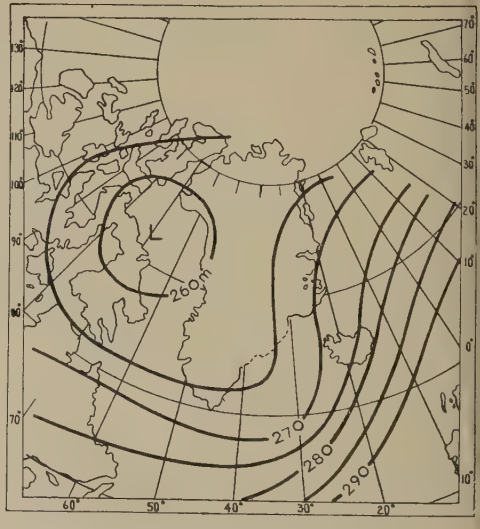

Figure 2. Mean Dec. 1953, 700 mb.

A study of the radiosonde ascents made at Station Centrale (71°N 41°W) (E.P.F. 1950) during December 1949 and January 1950 shows that in almost all cases there was a rapid temperature rise in the lowest 100 m above the surface, and that there was no appreciable temperature change in the next 500 m. The ascents indicate that there is a relationship between the temperature rise ΔT through this low inversion and the cloudiness and wind speed. A good empirical relationship was given by the equation

$$\Delta T = 15\,(8\text{-}n)/8 + 0{\cdot}31\,(30\text{-}v)$$

where n is the cloudiness measured in oktas and v the wind velocity in knots, over the range $v = 4$ to $v = 30$.

Of the 36 soundings made at Station Centrale during this period when the wind speed was less than 34 kt (gale force) two made during moderate snow and four made in heavy drifting snow have been excluded from the analysis. In the remaining 30 cases the standard error of ΔT estimated by using the above equation was 3·2°C.

It has been assumed that at Northice, which is 800 km further north but similarly situated on the almost horizontal part of the Greenland ice sheet, the temperature profile in the lowest 500 m is similar to that at Station Centrale. For each synoptic observation at Northice the temperature T_{700} of the 700 mb surface has been estimated, by use of the above equation, from the recorded values of screen temperature, wind speed and cloud amount. Except when there was a change of air mass the variation from one observation to another was small and amounted normally to two or three degrees. The values of T_{700} for 03 hr* and 15 hr, the hours of the 700 mb charts, were estimated by interpolation between the values for the nearest synoptic hours (00, 06, 12 and 18 hr). The mean temperature $\overline{T}$ of the layer of air between screen level and the 700 mb surface is then estimated, and the altitude of the 700 mb surface is read from the table described in § 4 of the article on midsummer (Hamilton 1958) for the measured station level pressure and the estimated value of $\overline{T}$.

The equation may not be strictly applicable at Northice, but it seems likely that the standard error in determining $\overline{T}$ at Northice is probably not greatly different from that at Station Centrale, i.e. 3·2°C. The mean value of $\overline{T}$ at Northice in the midwinter period was about 250°A, and the mean pressure 735 mb, with extreme values of 758 and 718 mb, so the height of the 700 mb surface above the barometer was always between the limits of 100 m and 600 m. The error in H_{700}, arising from an error of 1°C in $\overline{T}$, is about 4 m.

In addition to this random error in H_{700}, there is a constant error due to the inaccuracy in the determination, by geodetic methods, of the altitude of the station. The ' probable error ' of the geodetic measurement has been estimated to be about 7 m.

The value of H_{700} at Northice in geopotential metres has been plotted on the 700 mb charts with an estimated standard error of about 10 m.

4. Frequency and examples of frontal activity

During December 1952 the 17 disturbances listed in Table 1 affected the weather at either or both of the two stations. Three of these were weak (those of 7, 9 and 31 December) and were not followed on the 700 mb charts, and the history of the fronts from the north could not be established owing to the lack of upper-air stations there.

Of the 17 disturbances, 14 consisted of the passage of a front or trough; in the 3 remaining cases a frontal wave passed near the stations. In 9 cases there was snowfall, in all cases light, at Northice and in only 2 of them was there any snowfall at Britannia Sø : in each case only a trace was measured.

During the month the airflow was generally SWly and fronts and depressions swept across Greenland. Fig. 3 shows a warm-sector depression passing south of the stations on 29 December. At 00 hr 29th at Northice a trace of Cc was seen on the horizon to the south.

* All times are GMT.

Cloud increased to 8/8 Ac by midday and by 15 hr there was light snow but soon afterwards the cloud cleared and the surface wind veered 30°. The pressure minimum was at 09 hr. At Britannia Sø cloud increased in the afternoon to 7/8 AcAs at 18 hr : it had been calm but at 19 hr the NWly wind set in and continued, with clearing skies, at gale force. The pressure minimum was at 14 hr. The time differences between the clearances and the pressure minima at the two stations were thus about five hours, indicating a speed of travel of about 40 kt in good agreement with the warm-air speed.

Fig. 4, the 700 mb chart for 15 hr 8th, is typical of a warm sector of an occluded depression crossing Greenland. The warm front passed Northice around midnight 7/8 December with 8/8 As and light snow : the pressure minimum was at 23 hr 7th and the wind veered 30° : the cloud was broken to 4/8 at 06 hr. The cloud increased again to 8/8 As at 12 hr on the approach of the cold front, there was no snowfall, and by 15 hr there was only 3/8 Cc and later clear skies. At Britannia Sø it was calm and clear early on 8th : the warm front gave 8/8 As and light snow at 12 hr 8th, but the surface air remained calm without change of temperature. The cold front passed at 17 hr with clearing skies and onset of NWly wind : the temperature rose 8°C and the relative humidity fell 50 per cent.

These two examples are typical of the disturbances which affected N. Greenland in the winter. They caused only light precipitation at Northice and generally no precipitation at Britannia Sø, where a layer of upper cloud was the only evidence of the frontal passage. Warm fronts did not normally disturb the stagnant pool of cold air at Britannia Sø, but cold fronts caused the onset of the NWly wind.

Fronts from the north caused only light snowfall at either or both stations.

Table 1 also lists 18 disturbances during January 1953, almost equal to the number for December. The characteristics of frontal passage were similar, except that of 1 January (Fig. 5) when a warm front passed Britannia Sø and caused the onset of the katabatic wind : presumably on this occasion the gradient wind was too strong to permit the cold pool to persist. On this occasion the temperature rose from — 23°C to 0°C and the relative humidity (with respect to water) fell from 98 per cent. to 58 per cent.

The only considerable snowfall of the two months occurred during the period 25/28 January. The charts for 25, 26 and 27 January (Fig. 6, 7 and 8) show a deep depression over S. Greenland. There was little snowfall on passage of the fronts in the evening of 25th, but there was continuous snowfall at both stations on 26th due presumably to convergence and orographic uplift in the warm moist air from the east. Pressure fell 16 mb at

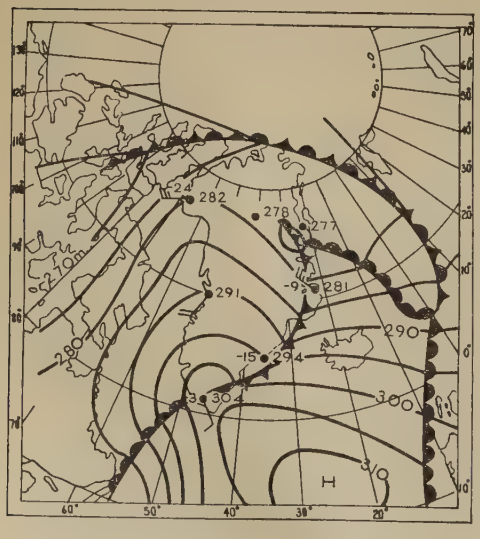

Figure 3. 1500 29 Dec. 1952, 700 mb.

Figure 4. 1500 8 Dec. 1952, 700 mb.

ritannia Sø during 26th, but at Northice the fall was less than 2 mb. A warmer air mass om the east passed Britannia Sø at 02 hr 27th with a marked temperature rise : this front eached Northice 16 hr 27th and snowfall ceased after its passage. There was a marked ressure rise amounting to 26 mb at Britannia Sø and 17 mb at Northice during the period 6 hr 27th to 06 hr 28th. The snowfall continued at Britannia Sø until the passage of a old front 00 hr 28th : this front caused a period of snowfall around 12 hr 28th at Northice.

During this period the measured snowfall at Britannia Sø amounted to 3·9 mm. The totals for December 1952 and January 1953 were Tr and 6·2 mm. No measurements f snowfall were made at this time at Northice but there is no doubt but that the total nowfall at this time was much larger than the sum of all the other falls of snow during he two months.

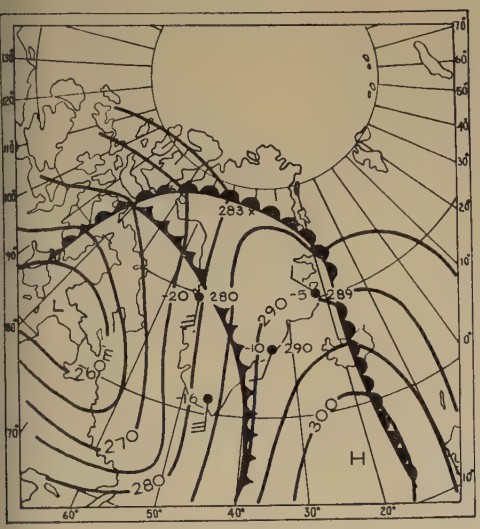

Figure 5. 1500 1 Jan. 1953, 700 mb.

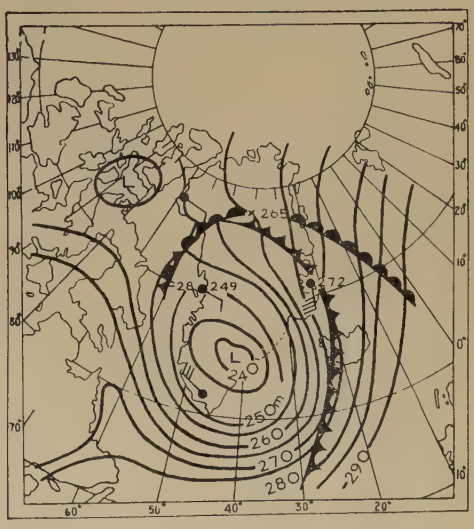

Figure 6. 1500 25 Jan. 1953, 700 mb.

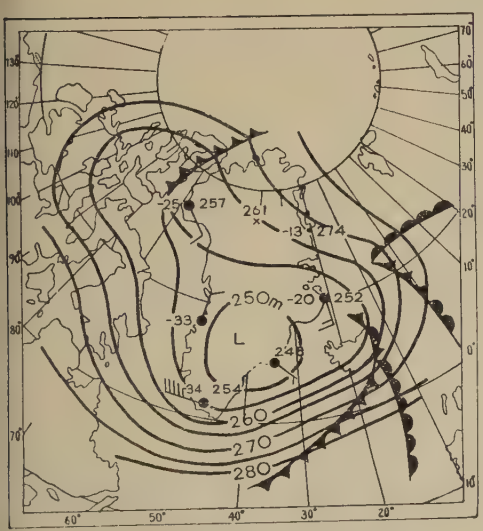

Figure 7. 1500 26 Jan. 1953, 700 mb.

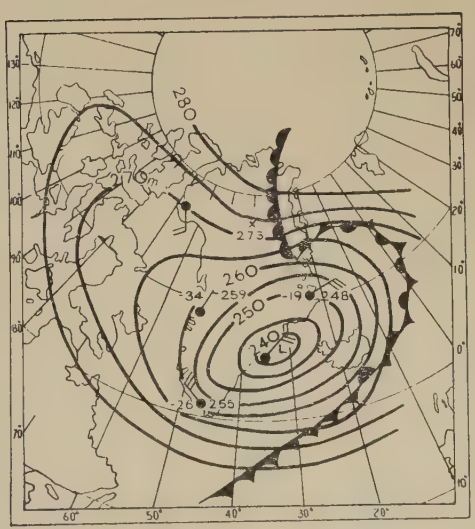

Figure 8. 1500 27 Jan. 1953, 700 mb.

TABLE 1. WEATHER ASSOCIATED WITH PRESSURE SYSTEMS 1 DECEMBER 1952 TO 31 JANUARY 1953

		Northice			Britannia Sø		
Dec. 1952	Frontal system at 700 mb ("N" = Northice B.S. = Britannia Sø)	Pressure minimum (GMT/date)	Precipitation (mm)	Weather (deg, kt, GMT)	Pressure minimum (GMT/date)	Precipitation (mm)	Weather (deg, kt, GMT)
1	Cold occlusion from SW : tip of warm sector just S of "N"	11/1	Light	Change of wind : air-mass temp. fall. Light snow	—	Nil	No cloud, but NWly wind ceas 16
4	Wave from WSW passed near to but S of "N" and B.S.	06/4	Nil	Wind dropped to light as wave approached. 8/8 thin As	11/4	Nil	Onset NWly wir at 10. 8/8 As, broke afternoon
6	Wave from WSW passed near to but S of "N" and B.S.	09/6	Nil	8/8 AcAs	18/6	Nil	Bank Ac 15. Wir dropped to light Ely then NWly again 01, 7th
7	Weak warm front from SW	—	Nil	8/8 Cs at 00	—	Nil	8/8 As at 21
7-8	Warm front from SW	23/7	Tr.	8/8 As. Light snow : temp. rise	—	Tr.	8/8 As with light snow at 12, 8th
8	Cold front from WSW	—	Nil	8/8 As then clearance at 15	—	Nil	Clearance with onset NWly wind at 17
9	Weak cold front from WSW	—	Nil	8/8 Cs	—	Nil	No cloud. Onset NWly wind at 15
10	Weak front from the N	18/10	Tr.	3/8 As with ice particles at 21	06/10	Tr.	8/8 As. Light snow 18
12	Weak front from NW	18/12	Tr.	2/8 Ac with light snow	—	Nil	Fine
14-15	Trough from W	19/14	Tr.	4/8 As with int. light snow. No temp. change	—	Nil	3/8 Ci 09, 15th
18	Warm occlusion from SW	21/18	Tr.	8/8 As : Light snow 23, temp. rise	—	Nil	Fine
22-23	Q.S. occlusion to W of "N"	15/22	Tr.	8/8 As with light snow at 06 and 18, 22nd, and 8/8 As 06, 23rd with hoar frost	—	Nil	Fine
23–26	Slow-moving cold occlusion from W	18/23	Nil	8/8 As with temp. fall	15/26	Nil	7/8 Ac 18, 26th after onset NWly wind
27-28	Warm front from SW	12/27 Fall steadied	Nil	8/8 Cs, temp. rise	—	Nil	8/8 Cs 06, 28th
27-28	Cold front from SW	—	Tr.	8/8 As with light snow, temp. fall	—	Nil	8/8 AcAs 18, 28th
29	Wave from SW passed S of "N"	09/29	Tr.	8/8 Ac with light snow 15. Winds light then strong	14/29	Nil	8/8 AcAs evening with onset NWly wind at 19
31	Weak warm front from SW	—	Nil	8/8 Cs; temp. rise	—	Nil	8/8 Cs at 06
Jan. 1953							
1	Warm front	06/1	Light	8/8 Cs with light snow	23/1 Fall steadied	Nil	Frontal passage at 23. 8/8 As with onset NWly wind. Temp. rose — 23° to 0°C

TABLE 1 (continued)

Dec. 1952	Frontal system at 700 mb (" N " = Northice B.S. = Britannia Sø)	Northice			Britannia Sø		
		Pressure minimum (GMT/date)	Precipitation (mm)	Weather (deg, kt, GMT)	Pressure minimum (GMT/date)	Precipitation (mm)	Weather (deg, kt, GMT)
3	Waves on cold front	—	Nil	Cold front passed 00, and warm front passed 12. Cloud and temp. changes	—	Nil	Variable cloud
4	Cold front became stationary between " N " and B.S. and frontolysed	09/4	Tr.	Frontal passage 09. 8/8 As and snowfall followed by clearance	—	Nil	Fine
7	Wave moving NE passed far to S of " N " closer to B.S.	—	Nil	1/8 As on S horizon at 09. Wind fell to calm afternoon	—	Tr.	8/8 As with snowfall
9	Occlusion or trough from N	—	Tr.	Frontal passage 17. As and snowfall.	18/9	Nil	8/8 As, clearing after 18
10	Cold front from the E	—	Nil	Fine	—	0·1	8/8 As with snow clearing afternoon
14	Warm front	12/14 Fall checked	Nil	Prelim. front 12, main front 18. 8/8 Cs and As marked temp. rise	—	Nil	Fine and calm
15	Cold front	18/15	Slight	8/8 Ns with cont. light light snow 12-17, clearing 18. Calm	18/5 Fall checked	0·5	8/8 Ns with snow 12. Clear by 18, NWly wind at 20
16	Wave moving N passed passed E of B.S.	—	Nil	8/8 Cs evening	06/16	0·5	8/8 As with snow and drift
20	Front or trough from N	—	Nil	Nil	—	Tr.	8/8 As with snow midday clearing evening
21-22	Warm front from SW	21/21	Nil	Marked temp. rise 21, 21st	09/22	Nil	Fine and calm
23	Warm front from SSE	—	Nil	7/8 As 18 broke 21 with marked temp. rise and Sly wind	—	Nil	7/8 As at 12
24	Cold front	—	Nil	Cloud cleared 06, 24th followed by temp. fall and wind veer	—	0·6	8/8 As with snowfall 18 onwards. Onset NWly wind 03, 25th
25	Occluding fronts	—	Light	7/8 As from 12, with snowfall 18 and SSE wind	—	0·1	8/8 As 18, snow later
26	Low centred S Greenland. Convergence and orographic air flow	—	Mod.	Continuous snow	—	2·0	Continuous snow
27	Warm front from E	—	Mod.	Frontal passage 16. Snowfall ceased, high temp.	02/27	1·1	Frontal passage 02. Continuous snow : marked temp. rise
28	Cold front	—	Light	Frontal passage 12. Snowfall, temp. fall clearance later	—	0·7	Frontal passage 00. Cessation of snowfall, temp. fall
30	Front from N	00/30	Tr.	Frontal passage 00. Snowfall; clearance of cloud	—	Tr.	Frontal passage 09. Cloud clearance, fall in temp.

5. The surface wind at Northice and Britannia Sø

(a) *Northice*

As the general direction of airflow at 700 mb was SWly, the prevailing wind (i.e. the wind 10 m above the surface) would be expected to be SSWly if there were no slope causing katabatic flow. The Climatological Tables show that the prevailing surface wind was westerly. It should be emphasized that Northice was on the open surface of the inland ice and there is no question of local funnelling of surface wind. This surface wind must be a katabatic wind, influenced by the general airflow above it. There were no occasions during the period on which there was no pressure gradient at 700 mb at Northice, but there were some occasions when there was only a small pressure gradient in a ridge of high pressure. The surface winds (in degrees and knots) on these occasions were as shown in Table 2.

TABLE 2. Surface wind at Northice

GMT	00		03		06		09		12		15		18		21	
	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt
15 Dec.													280	14	280	14
16 Dec.	280	14			270	15			270	17	280	17	270	15		
17 Dec.									270	13	270	12	280	13		
19 Jan.									280	08	280	10	280	13		
29 Jan.									280	17	290	17	290	17	290	17
30 Jan.	290	18			290	14			280	10	280	10	280	10	280	13

It appears that when there is little pressure gradient at the 700 mb level there is a katabatic wind of about 280° 14 kt. This is the same value as the midsummer katabatic wind. It is veered 40° on the direction of maximum slope (240° – the upward slope in this direction is about 9′). At Station Centrale the direction of the katabatic wind was found by Bedel (1954) to be veered 30° on the direction of maximum slope.

(b) *Britannia Sø*

It is not possible to determine accurately the direction of the m.s.l. pressure gradient at Britannia Sø as the only pressure measurements available are on the N-S-running east coast. The direction of the wind is strongly influenced by the topography which funnels the wind through the valley in which the station was situated : the contours of the steep hillside to the north of the station run in a direction approximately 290°–110°. The Climatological Tables (Table 2 (b) and 3 (b)) show that the predominating direction of strong winds was around 300°. Winds in excess of 6 kt from any other direction were rare though calms and very light winds were frequent. The records show that normally the onset of the NWly wind was sudden and it reached its highest velocity almost at once. It persisted usually for a few days, gradually decreasing, and then would drop suddenly from a speed of about 12 kt to almost calm. The calm conditions then persisted for a period until the next onset of NWly wind.

The times of onset and cessation of the NWly wind and the duration of the spells of NWly wind and calm conditions during December 1952 and January 1953 are shown in Table 3. (The period is extended into February so as to include an equal number of windy and calm periods). During this period there were 29 changes of wind, so the average duration of each spell was about 50 hr, but the average is reduced to this value by the frequent changes which took place during some periods of apparently critical stability, e.g. there were 10 changes in the eight-day period 1 December to 9 December. The total duration of calms (801 hr) was slightly greater than the duration of windy periods (737 hr).

TABLE 3. PERIODS OF WIND AND CALM, DECEMBER 1952 – JANUARY 1953, BRITANNIA SØ

Wind	Duration (GMT/date) From	To	Duration (hr)	Initial windspeed (kt)	Highest windspeed (kt)	Temp. change (°C)	R.H. change per cent	Vapour pressure before wind (mb)	Vapour pressure after change (mb)	Frontal change
NW	04/1	20/3	64	30	46	+ 12	− 55	0·7	1·1	
Calm	20/3	10/4	14			− 7	+ 40	0·6	0·8	
NW	10/4	07/5	21	25	42	+ 4	− 40	0·8	0·6	F
Calm	07/5	21/5	14			− 5	+ 24	0·6	0·7	
NW	21/5	15/6	18	26	26	+ 8	− 40	0·7	0·7	
Calm	15/6	01/7	10			− 2	+ 22	0·8	0·9	
NW	01/7	09/7	8	30	30	+ 9	− 32	0·9	1·0	F
Calm	09/7	17/8	32			− 4	+ 30	1·0	1·0	
NW	17/8	02/9	9	14	18	+ 8	− 48	0·9	1·3	F
Calm	02/9	15/9	13			− 8	+ 36	1·3	1·0	
NW	15/9	17/13	98	24	30	+ 9	− 41	0·9	1·1	F
Calm	17/13	03/16	58			− 5	+ 26	0·7	0·7	
NW	03/16	03/18	58	26	30	+ 8	− 36	0·6	0·8	
Calm	13/18	10/26	189			− 3	+ 14	0·9	0·8	
NW	10/26	12/27	26	36	43	+ 8	− 33	0·6	0·8	F
Calm	12/27	19/29	55			− 6	+ 21	0·8	0·7	
NW	19/29	00/31	29	34	36	+ 13	− 40	0·9	2·0	F
Calm	00/31	23/1	47			− 7	+ 21	1·6	1·3	
NW	23/1	20/2	21	22	26	+ 23	− 38	1·0	2·8	F
Calm	20/2	00/4	28			− 8	+ 24	2·0	1·8	
NW	00/4	06/8	102	24	48	+ 6	− 34	1·3	1·3	
Calm	06/8	20/8	14			− 4	+ 17	1·6	1·2	
NW	20/8	08/10	36	17	36	+ 11	− 32	1·1	1·7	F
Calm	08/10	20/15	132			− 3	+ 20	1·4	1·2	
NW	20/15	14/21	138	34	44	+ 8	− 10	0·6	1·2	F
Calm	14/21	09/26	115			− 7	+ 19	0·9	0·7	
NE	09/26	12/28	51		54	Blizzard				F
NW	12/28	18/30	54	30	36					
Calm	18/30	02/3	80			− 4	+ 8	1·1	0·7	

F in the last column indicates that the wind change marked the passage of a front listed in Table 1.

The relative humidity is measured with respect to water.

In the cases of Calm, the changes in temperature, relative humidity and vapour pressure are those which took place in four hours after the cessation of the NW wind.

The onset of the NWly wind is what Hobbs (1945) refers to as a surge of the centrifugal wind. It is clear from the Tables 1 and 3 that the onset of the NWly wind can almost always be related to a disturbance on the synoptic chart and usually marks the passage of a front.

During clear weather in the dark winter period there is a constant loss of heat by radiation from the surface, and the air in contact with the surface is cooled. On a slope this causes the downhill katabatic wind to develop, but on a horizontal surface there is a tendency for a stable inversion to form and, in the absence of any strong pressure gradient the surface wind, at the bottom of the inversion, becomes light or calm. In sheltered places, therefore, a cold stagnant pool of air forms whenever the pressure gradient becomes light. Although strong winds frequently poured through it, the valley containing Britannia Sø was afforded a certain amount of shelter by the higher land of Dronning Louise Land and calm conditions there often occurred while a strong katabatic wind was pouring off the inland ice. On 29 October 1952 the writer experienced a wind estimated at 50 kt at the foot of the steep part of the Suzanne Brae 15 km NW of Britannia Sø,* but after proceeding 2 km down the Britannia Gletscher the wind was found to be calm; at Britannia Sø it had been calm all the time. Again, strong winds often funnelled through the Britannia

* Maps of the area are published in *Geogr. J.*, **122**, 1956.

TABLE 4. LOWER WINDS AT BRITANNIA SØ

	Nov. 1952				January 1953						February 1953					
GMT/date	03/8		20/11		18/12		17/22		14/31		14/15		14/16		14/17	
Altitude (km)	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt
S	Calm		Calm		Calm		Calm		Calm		Calm		Calm		290	24
0·5	310	01	260	02	280	05	120	02	Calm		330	06	100	04	310	30
1	320	05	230	02	300	07	340	03	090	01	350	13	280	15	330	20
2	310	35	310	14	270	17					230	18	330	20	330	50

	February 1953										October 1953					
GMT/date	14/18		14/20		13/21		13/22		15/15		15/16		14/17		13/20	
Altitude	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt
S	310	14	300	22	300	18	300	20	320	01	300	06	050	01	220	01
0·5	310	21	280	26	270	20	280	28	110	02	110	03	090	03	230	01
1	340	22	330	11	250	06	270	25	190	03	130	10	330	01	290	02
2	320	45	340	12	330	15	350	25			200	23	190	14	180	17

	Oct. 1953		Jan. 1954		February 1954					
GMT/date	14/22		20/12		16/20		16/27		13/28	
Altitude	deg	kt	deg	kt	deg	kt	deg	kt	deg	kt
S	240	16	290	01	090	01	090	04	130	05
0·5	280	11	Calm		280	02	260	05	320	23
1	310	06	090	01	160	08	290	06	330	30
2	150	06					320	22		

Sø valley when the wind was light in a neighbouring locality. During the period 16 to 21 November 1952 a party crossing Strostrømmen from Dronning Louise Land to Germania Land reported light winds while strong winds persisted at Britannia Sø.

Table 4 shows the wind in the lower layers measured by the few pilot balloons which were released during the winter period 15 October to 28 February. The winds are the surface winds (corrected to 10 m), and the mean wind in layers 300 m thick centred at the height 0·5, 1·0 and 2·0 km above station level. They show in almost all cases that when the surface wind was light the wind at 1 km was also light, exceeding 10 kt in only two cases, except on 28 February 1954 which was an exceptional case with a light wind at the surface and strong winds aloft. On this occasion the surface wind had changed to 300° 28 kt half an hour after release of the balloon. The cold stagnant pool formed only when the pressure gradient was small. In almost all such cases the wind at 2 km was comparatively strong, showing that the depth of the cold pool was between 1 and 2 km.

When the surface wind was strong NWly the 1 km wind was normally westerly, but often lighter than the surface wind. Thus the NWly wind is not a true katabatic wind, but nor is it a true gradient wind, or else the velocity would increase with height. It appears to be a katabatic wind which is stronger at the surface than at 1 km but which blows only when there is sufficient gradient to prevent the formation of the stagnant cold pool. As the gradient decreased the wind decreased to a value around 12 kt but was then unable to maintain the katabatic flow which often ceased almost abruptly and the wind dropped to calm.

6. TEMPERATURE

(*a*) *Northice*

The screen temperature at Northice varied considerably with the wind speed and cloudiness. If the equation given in § 3 is applicable, then the difference between the air-mass temperature (i.e., the estimated temperature above the inversion) and the screen temperature would be 0°C for 8/8 cloud and 30 kt wind, 9°C for 8/8 cloud and no wind,

5°C for no cloud and 30kt wind and 24°C for no cloud and no wind. Changes of screen mperature of 15°C to 20°C can be expected, and did occur, over quite a short period the wind changes by about 10 kt and the sky changes from overcast to clear or vice versa.

Marked changes of temperature due to change of air mass also occurred. In the months nder discussion the highest screen temperature was — 18°C at 06 hr 1 January when warm r from the south (Fig. 5) spread over Greenland accompanied by light snowfall (estimated ir-mass temperature — 9°C). Temperature fell to — 28°C (estimated air-mass tempera- ure — 23°C) at 12 hr 2 January after passage of the cold front. The lowest temperature f the 1952/53 winter, — 57°C, occurred at 06 hr 19 January with a clear sky and wind peed of 10 kt (estimated air-mass temperature — 36°C). This day was in the middle of a ng spell, from 12 hr 17 January to 18 hr 21 January, of clear skies during which the emperature remained below — 50°C and the wind was generally between 10 and 18 kt. The spell was broken by a weak warm front which passed about 19 hr 21 January and was ccompanied by a small amount of cirriform cloud, a pressure minimum and a wind veer, ut the screen temperature rose from — 56°C at 15 hr with 1/8 Cs, and wind 240° 17 kt estimated air-mass temperature — 39°C) to — 41°C at 21 hr with 1/8 Cs and wind 270° 2 kt (estimated air-mas temperature — 22°C). There was normally no marked progressive emperature change during a fine spell when heat was being continuously lost from the now surface by radiation. In the case just quoted it was overcast with heavy drifting snow nd a temperature of — 41°C at 21 hr 16 January. The cloud cleared around midnight nd the temperature was — 48°C by 06 hr and — 53°C by 21 hr 17 January. Thereafter here was no progressive fall of temperature until the warm front passed on the evening f 21 January.

In the corresponding months of the following winter the highest temperature was — 20°C, but the lowest temperature was — 66°C, which occurred at 12 hr 9 January 1954 vith a wind of only 4 kt and a clear sky.

(*b*) *Britannia Sø*

As at Northice the temperature must be discussed in relation to the wind. The normal ccurrence was for the temperature to rise with the onset of wind and to fall when the wind ell to calm. This is mainly the result of the breaking down by the wind of the inversion n the cold pool and of the formation of the inversion in still air, but as in most cases the nset of wind was the result of frontal passage the rise of temperature depended also on he general temperature of the new air mass. As at Northice a remarkable change of air emperature occurred with the passage of the warm front about 23 hr 1 January 1953 (Fig. 5). The temperature rose from — 23°C at 18 hr to 0°C at 00 hr 2 January, the highest emperature during this midwinter period (on 5 February 1953 in a similar case the tem- perature at Britannia Sø rose temporarily to + 3°C). So great was the temperature change brought about by this warm front that from 00 hr to 21 hr 1 January 1953, while the front was moving from Northice to Britannia Sø, the screen temperature at Northice was actually higher than it was at Britannia Sø. This must be a very rare occurrence. The thermograms and hygrograms for the week including this period (Figs. 9 and 10) show three abrupt changes of temperature and humidity corresponding to onsets of the NWly wind (see also Table 3). Unfortunately the drum of the thermograph ceased to rotate on New Year's day (though the clock continued to work) and was not put into working order again until 09 hr 2 January so the trace is broken during this period. The dotted line shows the probable form of the trace based on the thermometer readings, the maximum thermo- meter readings and the corresponding hygrograph trace. On the occasion just quoted the temperature change was greater than normal, but the thermograph and hygrograph records show abrupt changes of temperature and humidity on the onset of the NWly wind on almost every occasion. The change of temperature and relative humidity (with respect to water) on the occasions of wind change are shown in Table 3. It will be seen that normally the temperature rose by about 8°C and the relative humidity (with respect to water) fell by about 30 per cent. The traces of the autographic instruments show that the change is very

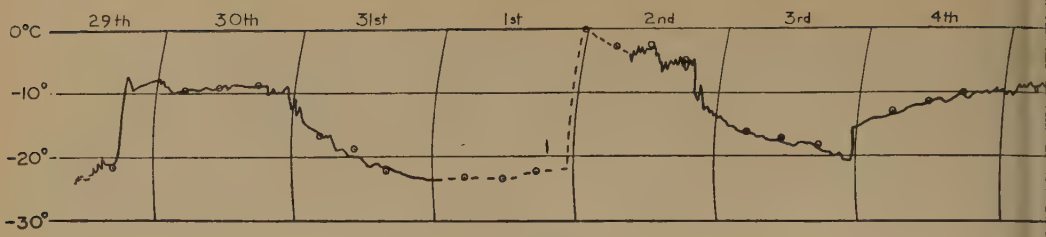

Figure 9.

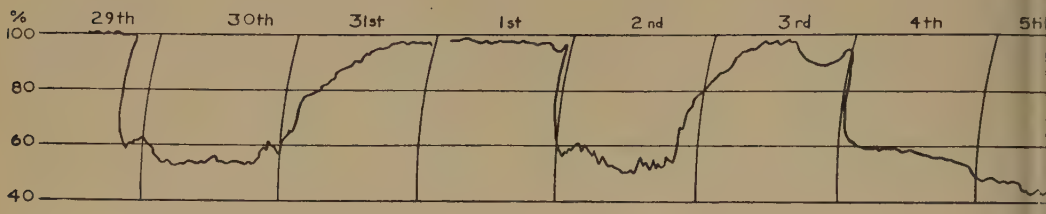

Figure 10.

abrupt. The abruptness of the temperature and humidity change on the onset of the NWly wind is not surprising, as the wind change is so sudden and great. The rapid fall of temperature and rise of relative humidity on the cessation of wind is more remarkable. Examples can be seen on the traces (Figs. 9 and 10) at 00 hr 31 December and 20 hr 2 January. The changes of temperature and humidity in the first four hours after the cessation of the NWly wind, as read from the traces of the autographic instruments, are shown also in Table 3. Part of the rapid fall may be attributed to lack of ventilation of the screen but experiments by the author (1956) showed that a thermometer exposed on the top of the screen read only about 1·5°C below the air temperature measured by a whirling thermometer. On two occasions the rapid initial fall of temperature was confirmed by hourly readings of whirling thermometer. On 16 November 1952 the temperature at 12 hr was −6°C with a wind 310° 14 kt, the wind ceased almost immediately afterwards, and the temperature fell to − 9°C at 13 hr, − 13°C at 14 hr and −14·5°C at 15 hr, a fall of 8·5°C in about three hours : thereafter the fall was slower. Again, on 12 February 1953 the temperature fell from − 25·5°C at 14 hr to − 33°C at 18 hr, a fall of 7·5°C in four hours. There was normally little difference between the water vapour pressure before the cessation of wind and that measured afterwards : i.e., in the former case it changed from 1·4 mb at 06 hr (wind 310° 30 kt) to 1·3 mb at 18 hr and in the latter case the change was from 0·4 mb at 06 hr to 0·3 mb at 18 hr.

The vapour pressures measured shortly before and shortly after each wind change are shown in Table 3. It will be seen that though there were sometimes marked changes of vapour pressure on the onset of the NWly wind, when this marks the passage of a front there was no appreciable change of vapour pressure when the NWly wind ceased. The table shows too, that there was little change of vapour pressure during the periods of calm except in the case of the long period 10 January to 15 January. The highest vapour pressure measured was 4·1 mb at 06 hr 27 January at a temperature of − 4°C during heavy drift snow (perhaps falling snow) on the arrival from the coast of maritime air from the N. Atlantic (Fig. 8). Thereafter the vapour pressure and temperature fell gradually as the wind backed slowly to the NW.

While the initial fall of temperature was normally rapid, the rate of fall soon decreased and a fairly steady temperature had usually been reached twenty-four hours after the cessation of the NWly wind. Table 5 shows the temperature 6, 24, 48, 72 . . . . hours after the cessation of the NWly wind for those periods of calm shown in Table 3 which exceeded three days.

TABLE 5. CHANGE OF TEMPERATURE DURING PERIODS OF CALM – BRITANNIA SØ

GMT/date of cessation of wind	Temp. before cessation of wind	hours							
December	(°C)	6	24	48	72	96	120	144	148
17/13	− 14	− 19	− 27	− 28					
13/18	− 21	− 24	− 25	− 28	− 27	− 29	− 29	− 29	− 30
12/27	− 16	− 24	− 26	− 24					
January									
08/10	− 14	− 19	− 28	− 29	− 29	− 30	− 27		
14/21	− 18	− 26	− 28	− 27	− 27	− 29			
18/30	− 19	− 26	− 37	− 36					

The thermogram reproduced as Fig. 9 shows the general shape of the cooling curve.

No equipment was available for making temperature soundings at Britannia Sø, so no measurement could be made of the temperature gradient in the cold pool or of its depth. In Nov. 1952 when the wind was variable but mainly light, temperature and aneroid readings were taken at several heights on the steep hillside to the north of the station : these showed a slight temperature fall of about 1°C in 150 m, but this is no larger than the normal hour-to-hour fluctuation of screen temperature in calm conditions.

On another ascent of the steep hillside between 12 hr and 13 hr on 15 February (when the sun was still below the horizon at midday) in calm conditions, with screen temperature − 37°C, the temperature increased with height at a rate of 2°C per 100 m to − 31°C at a height of 270 m above lake level, and then was isothermal for a further 100 m. A pilot-balloon ascent made immediately afterwards showed that up to a height of 250 m the wind was only 3 kt from WNW but then increased in the next 100 m to 330° 11 kt. Cloud increased subsequently and though the wind remained light the screen temperature rose to between − 29°C and − 31°C during the night and then fell again next day, when the cloud cleared, to − 37°C. It appears that on this occasion the inversion was about 250 m deep, with a temperature difference of 6°C between its base and top.

Arnold, one of the Expedition's surveyors, determined the temperature gradient in the air layer 2 to 7 m above the frozen lake surface, on a few occasions, by measuring the optical index of refraction, using the same method as used by the author (1957) at Northice. As the surface of the lake may not have been truly horizontal there is a little uncertainty in the absolute calibration, but changes of temperature gradient are accurately measured by this method. The measurements showed that during a windy period the temperature gradient was about isothermal, but in two cases of calm clear conditions the gradients were + 7°C and + 13°C per 100 m. Such gradients near the surface are to be expected when there is a mean gradient of 2°C per 100 m in the lowest 250 m.

It appears, therefore, that as soon as the wind ceases the lowest layers of the air are cooled quickly by contact with the radiating surface. The heat is transferred downwards by turbulence, the gradient being greatest near the surface where the eddy conductivity is smallest, and decreases upwards where there is greater turbulence, and the eddy conductivity is greater. Radiational exchange between different atmospheric layers presumably enters also into the heat balance. The cooling at first occurs in the lowest layers and then extends upwards, until a steady state is reached when the heat lost by radiation is replaced by heat transferred downwards from the air and from heat flowing upwards through the ground, or the lake ice. The temperature measurements at Britannia Sø lead to the conclusion that in the cold pool the temperature difference across the inversion is normally about 8°C and suggest that the depth of the pool may be about 300 m, above which the rate of change of temperature seems to be small.

7. Visibility. Drift snow

(a) *Northice*

In general the visibility was good except in so far as it was limited by drift snow. Light winds were infrequent, and radiation fogs and mists occurred only rarely.

During the period November 1952 – February 1953 there were only five occasions of visibility less than 1 km in which the reduction was caused neither by falling or drifting snow. At 18 hr 15 January a visibility of 180 m was recorded in calm conditions immediately after the passage of a cold front and cessation of snowfall : intermittent snow was reported later and the reduction of visibility may have been due to small snow particles which could not be easily distinguished in the dark. A visibility of 800 m was reported at 21 hr 27 January immediately after cessation of snowfall accompanying the warm front (Fig. 8) bringing high temperatures to Northice. Three more observations of fog (400-900 m) were made during the evening of 28th after the cessation of snow on the subsequent cold front. In these last four cases the wind speed was 2 to 3 kt.

There were, therefore, no cases of true radiation fog : in these few cases the fog was associated with the passage of a front and may have been due to precipitation which could not be distinguished in the dark.

All other cases of poor visibility were due to falling and drifting snow. Rollitt, one of the Expedition's meteorologists, made a careful record of the height and intensity of the drift snow. Table 6 gives the total number of observations during the period November 1952 – February 1953 for the given wind speeds, of drift snow (without falling snow) according to the following international code :

N = No drift snow
0 = Drift snow close to the ground (up to 20 cm)
2 = Drift snow near the ground (20-150 cm)
4 = Drift snow, light or moderate
6 = Drift snow, moderate or heavy
8 = Drift snow, heavy, (impossible to observe whether snow is falling or not)

It should be noted that during this period the anemometer was at a height of 2 m above the surface : the 10 m mast could not be erected until April 1953. In Table 6 (and in Table 8) the measured wind speed at 2 m is given as well as the wind at 10 m estimated according to Hellmann's formula.

TABLE 6. Variation of height and intensity of drift snow with wind speed

Wind speed at 2 m (kt)	Est. wind speed at 10 m (kt)	No. of occurrences of Code Figure						Total
		N	0	2	4	6	8	
0–6	0–8	107						107
7	9	27	4					31
8	10	32	12	1				45
9	12	18	29	2				49
10	13	14	19	29				62
11	14	6	14	29	4			53
12	15	5	8	34	3			50
13	17	9	5	16	37			67
14	18	1	1	4	32			38
15	19	3	2	9	37			51
16	21	3	0	2	13			18
17	22			1	31			32
18	23				12	7		19
19–20	24–26			1	18	10		29
21–23	27–30				9	14		23
24–31	31–40				1	7	1	9

It will be seen from Table 6 that the height and intensity of the drift snow increases rapidly as the wind speed (at 10 m) increases from about 12 kt to about 17 kt. The critical wind speeds for each code figure for drift snow are shown in Table 7 : the critical wind speed is here defined as the wind speed at which two adjacent code figures occurred equally frequently. These measurements are to a certain extent subjective, so the figures for each month are included to illustrate the degree of consistency from month to month.

TABLE 7. CRITICAL WIND SPEEDS (AT 10 M) FOR DRIFT SNOW (KT)

Month	Adjacent Code Figures			
	N – 0	0 – 2	2 – 4	4 – 6
November 52	13	15	18	29
December 52	10	13	17	24
January 53	11	14	16	24
February 53	11	13	16	27
4 months	11	13	16	27

It is difficult to account for the exceptional values in Table 6, in particular for the occasions of no drift when the wind was strong. Of the three cases of no drift with a wind speed of 21 kt, two occurred during the very warm conditions at 00 hr 1 January and 16 hr 25 January, but the third occurred at 00 hr 25 December in apparently unexceptional circumstances with a temperature of − 46°C after a spell of 3 days without drift snow which continued until the evening of 25 December when light or moderate drift snow was observed with a wind speed of 22 kt and an overcast of cirrostratus. Weather had been fairly fine over Greenland for some days, with only a weak occlusion which gave no precipitation at Northice, and it seems possible that the lack of drift snow with these moderate winds may be due to the fact that no fresh snow had fallen anywhere on the inland ice for several days.

The implication is that once the snow surface has been compacted it is not readily eroded by the wind, and drift snow may occur again only when a fresh supply of snow has fallen to windward of the station.

Although by definition drift snow of Code Figures 0 and 2, with maximum heights of drift of 20 and 150 cm, does not restrict the horizontal visibility the observations show that the visibility decreases with the wind speed at all speeds above 8 kt. The lowest wind speed at which a height of drift exceeding 150 cm was observed was 14 kt, so it must be concluded that in cases of lower wind speeds the visibility on occasions is restricted by smaller drift particles which have risen above the top of the main layer of drift snow.

Table 8 shows the variation of visibility with wind speed. For each wind speed the number of occurrences of each visibility code figure was counted and thus the median visibility was computed. To illustrate the degree of variation the upper and lower quartile visibilities are also given.

Liljequist (1957) has shown that on an antarctic snow field there is a linear relationship between log u, where u is the wind speed at 10 m measured in m/sec, and log V, where V is the visibility measured in metres. Over the visibility range 7 to 700 m approximately the relationship is

$$\log V = 8{\cdot}0 - 5 \log u$$

The values of V deduced from this equation are shown in the last column of Table 8. It will be seen that this relation is in good agreement with observation on the Greenland ice sheet over the range 50 m to 10 km.

TABLE 8. Variation of visibility with wind speed at Northice November 1952 – February 195

Wind speed at 2 m (kt)	Est. wind speed at 10 m (kt)	No. of Observations	Lower quartile	Visibility median	Upper quartile	Theoretical value
7	9	31	10 km	18 km	45 km	46 km
8	10	45	12	18	30	27
9	12	49	2·5	8	18	11
10	13	62	1·8	8	18	7
11	14	53	1·7	7	18	5
12	15	50	950 m	3	5	3·6
13	17	67	850	1·7	11	1·9
14	18	38	850	1·5	5	1·5
15	19	51	300	1·5	4	1·1
16	21	18	350	1·0	2	680 m
17	22	32	250	600 m	1·5	540
18	23	19	120	400	550 m	430
19–20	24–26	29	120	250	600	280
21–23	27–30	23	90	120	190	150
24–31	31–40	9	35	50	90	90

(*b*) *Britannia Sø*

Although there were many periods of clear calm weather, conditions which would be suitable for radiation fog in temperate latitudes, the visibility in calm conditions was always over 20 km except in the few cases when snow was falling.

At the temperatures which prevailed in the calm clear conditions, e.g. — 20°C to — 40°C, the water-vapour content of the air is very low, and even deposits of hoar frost were only small and did not occur until the air was saturated with respect to supercooled water (Hamilton 1956). It is not surprising, therefore, that in the free air, which probably has a low nuclei content, mists and fogs did not occur.

Owing to the roughness of the terrain, measurements of the height and intensity of drifting snow were more difficult to make than at Northice, but measurements of visibility were more accurate as geographical features were visible over a wide range. Drift snow was only infrequently observed at Britannia Sø. During the months of November and December 1952 there were no observations of drift, though the wind speed exceeded 21 kt on 35 per cent of occasions and exceeded 33 kt on 12 per cent of occasions. On 98 per cent of occasions the visibility exceeded 40 km and on all occasions exceeded 10 km. During these two months there were only 5 very light falls of snow, in each case amounting to only a 'Trace' and it can be presumed that no appreciable amount of snow fell in the neighbourhood. This absence of drift snow must be due to the lack of a source of supply.

During January drift snow was reported on three occasions.

(i) 21 hr 15 January on the onset of NWly wind after a period of snowfall, total 0·5 mm, during the day. (This was the first snowfall exceeding 0·1 mm during the winter). The wind speed was 34 kt and the visibility 3 km. The drift continued on the 16th when the wind speed was 24 kt and the visibility improved from 5 km to 15 km. During the evening there was further snowfall followed by a short period of light drift.

(ii) Low drift snow was reported at 18 hr 20 January with a wind speed of 44 kt and visibility 16 km after a light snowfall at midday.

(iii) After the heavy snowfall 26 to 28 January, drift snow continued throughout 29th; the visibility being 1·6 km with a wind speed of 36 kt at 09 hr and improving to 8 km with a wind speed of 26 kt at 21 hr. No drift was reported on 30th though the wind speed exceeded 20 kt until the afternoon.

During February there were five periods of drift snow. Of these three occurred after snowfall, the fourth was no more than a single observation of low drift with a visibility of over 40 km. The fifth period was 19 to 21 February, on the first day of which the sky was obscured and it could not be definitely determined whether snow was falling or not. At this time pressure was low over the N. Atlantic and a northerly airstream swept from the

Arctic Ocean over the NE part of the inland ice. Continuous snow was falling at midday on each of these days at Nord (82°N 18°W) and it seems likely that snow fell extensively on the ice sheet, providing a source from which the snow could drift to areas where snow did not actually fall.

During the following winter there was very much more snowfall at Britannia Sø, and drift snow was more frequently reported : nevertheless drift snow was not normally reported in strong winds except after a snowfall.

(c) *Discussion*

The contrast between the drift-snow conditions at Northice and at Britannia Sø are so remarkable that a short discussion is necessary. Some allowance must be made for the different sites. At Northice the observer was surrounded by a flat layer of snow for hundreds of kilometres in all directions : the wind was normally very steady and the air was probably very stably stratified. Measurements of visibility were made from observer's eye to approximately the same level. At Britannia Sø the observations were made on a rough moraine in a deeply dissected landscape and wind was normally turbulent. The visibility observations were made from the site of the meteorological instruments which were 9·5 m above lake level, normally towards the south side of the lake. It is possible that low drift blowing over the lake, but not over the moraine, might have been missed sometimes but drift snow of moderate intensity would certainly be observed. Lack of drift snow in the NWly wind and of the drifting of snow round objects was a conspicuous feature of the locality as was, indeed, the small amount of snow deposited on the land and glaciers for many kilometres around.

It has been establised that drift snow occurred at Britannia Sø only after a period of snowfall in the neighbourhood. On the other hand drift snow was a characteristic of the Suzanne Brae, the glacier 15 km to the NW which flows steeply from West to East and forms the lower part of the inland ice. It appears, therefore, that the surface air which, as a NWly wind, (but not a true katabatic wind) pours through the valley of Britannia Sø cannot be air which had flowed off the inland ice at surface level and has then been diverted down the Britannia Gletscher, or else it would bring its burden of drift snow with it – unless, of course, the drift snow is subsequently carried to higher levels by turbulence. It appears that the air which blows near the surface of Britannia Sø must have come from a higher level, at which there is no burden of drift snow, and then descended down the Britannia Gletscher. Drift snow occurs, therefore, only if there is a supply of snow available for lifting on the Britannia Gletscher or in its neighbourhood.

On the inland ice drift snow appears to be the normal accompaniment of wind, leading one to infer, perhaps, that it is actually extracted from the snow surface itself. While it seems likely that drift particles will cause erosion of the snow surface and give rise to further drift particles, the Britannia Sø observations indicate that the wind alone will not raise drift particles from a well compacted surface. That there is a net accumulation of snow at Northice in spite of almost continuous drift snow shows that the snowfall is greater than the net removal of snow by drifting. If there is a good supply of soft snow to the windward of Northice – perhaps near the crest of the ridge, or to the west of it – then the large amount of drift snow observed at Northice may be snow from this source drifting past the station without appreciable accumulation or erosion. Wind and drift snow observations from a number of stations on the inland ice are required to enable a study of drift snow to be made.

8. Review of midwinter weather 1952-53

The foregoing pages show that during the midwinter period December 1952 and January 1953 the weather in the area of N. Greenland under consideration was normally fine and with a pattern of surface winds which showed little change from day to day. The fine weather and the normal wind system was broken down by travelling cyclonic systems and their associated fronts causing snowfall which was mainly light in character.

In December 1952, of the 17 disturbances listed in Table 1, only 2 caused precipitation at both stations, 7 caused precipitation at Northice only and 8 caused no precipitation at all. No snowfall greater than ' Trace ' occurred at Britannia Sø.

In January 1953, 18 disturbances are listed, 6 of which caused precipitation at both stations, 3 caused precipitation at Northice only and 5 caused precipitation at Britannia Sø only, while 4 caused no precipitation at either station. Only 2 disturbances caused precipitation amounting to 1 mm or more at Britannia Sø and for the whole period of two months 80 per cent of the precipitation fell during the period of 24 to 28 January and was associated with one deep depression off the east coast. It was during this period that the only moderate snow falls at Northice occurred.

(*a*) *Comparison of midwinter 1952-53 with 1953-54*

It is necessary to follow up this detailed study of the weather of December 1952 and January 1953 by considering whether these months were fairly normal and may be regarded as representative.

The following statistics enable some comparison to be made between the two winters :

TABLE 9. COMPARISON OF FOUR MIDWINTER MONTHS

	Northice			
	December 1952	January 1953	December 1953	January 1954
Mean cloud amount (eighths)	3·1	3·3	2·3	1·9
Mean low cloud amount (eighths)	0·3	1·0	0·8	0·5
Mean wind speed (kt)	14·5	13·6	15·5	17·9
Percentage of winds (210-020 deg)	100	87	92	90
Percentage of winds (030-200 deg)	0	10	8	6
Mean station level pressure (mb)	744·3	736·3	729·5	737·9
	Britannia Sø			
	December 1952	January 1953	December 1953	January 1954
Mean cloud amount (eighths)	1·3	3·3	3·8	2·8
Mean low cloud amount (eighths)	0·0	1·1	2·0	0·8
Mean wind speed (kt)	12·5	16·4	15·9	13·0
Percentage of winds (210-020 deg)	55	61	73	60
Percentage of winds (030-200 deg)	13	14	18	26
Mean station level pressure (mb)	990·0	980·7	971·1	983·4
Number of days of precipitation (a) 0.1 mm or more	0	8	10	3
Number of days of precipitation (b) 1·0 mm or more	0	2	6	2
Total precipitation mm	Tr	6·2	18·0	8·1

An analysis of the synoptic charts for December 1953 and January 1954 was made and a table (not published) similar to Table 1 was drawn up. It showed that in December 1953 there were 13 disturbances of which 12 caused precipitation at Britannia Sø and 6 of these also caused precipitation at Northice. 6 of the disturbances caused more than 1 mm at Britannia Sø. In January 1954 there were 14 disturbances of which 2 caused precipitation at both stations. 4 caused precipitation at Britannia Sø only, 2 caused precipitation at Northice only, and 6 caused no precipitation at all. Only 2 caused more than 1 mm precipitation at Britannia Sø.

It appears that the two Januarys were similar at both stations but that the weather of December 1952 was better and that of December 1953 was worse than in either of the Januarys, which were fairly similar. The pressure differences are very marked. The mean

700 mb charts for the two Decembers (Figs. 1 and 2) show the different patterns of pressure and airflow at 3,000 m approximately and indicate why the precipitation was much less in December 1952, when there was a ridge of high pressure over Greenland and the airflow was westerly over N. Greenland, than in December 1953 when there was a trough of low pressure over Greenland and a southerly airflow. Figs. 11, 12, 13, 14 (reproduced from *Die Grosswetterlagen Mitteleuropas*) show the m.s.l. pressure anomalies for the four months. From these it will be seen that pressure was markedly high over N. Greenland in December 1952 and markedly low in December 1953. In the Januarys the pressure anomaly was only slightly positive. It may be presumed, therefore, that the two Januarys were fairly normal while the two Decembers were abnormal.

It is interesting to note the relatively small variation in the monthly number of disturbances affecting the area concerned, viz. 17, 18, 13, 14. The numbers for June 1953 and June 1954 were 19 and 13.

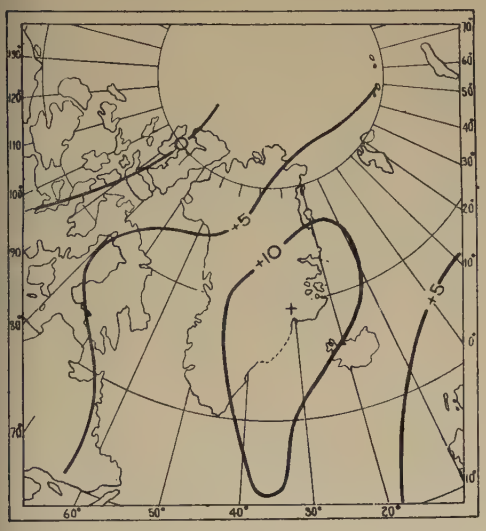

Figure 11. M.S.L. pressure anomaly, Dec. 1952.

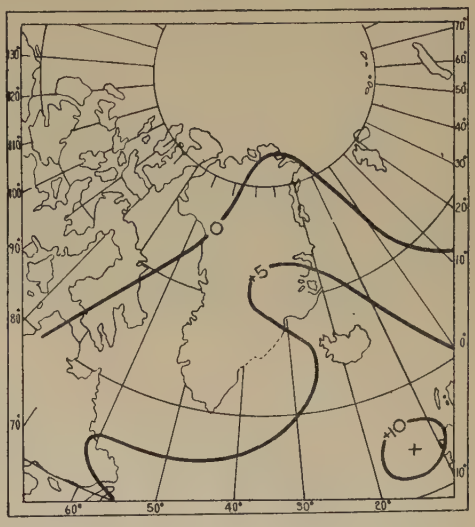

Figure 12. M.S.L. pressure anomaly, Jan. 1953.

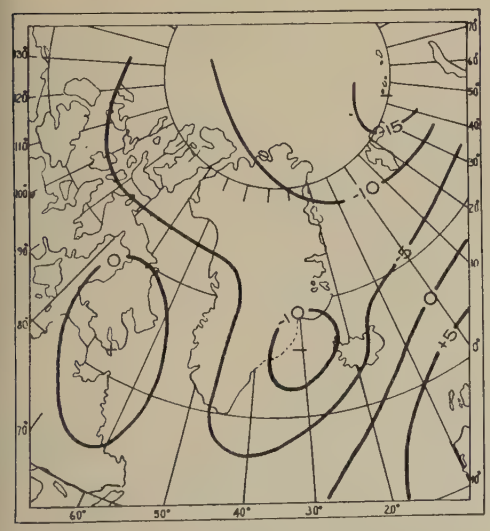

Figure 13. M.S.L. pressure anomaly, Dec. 1953.

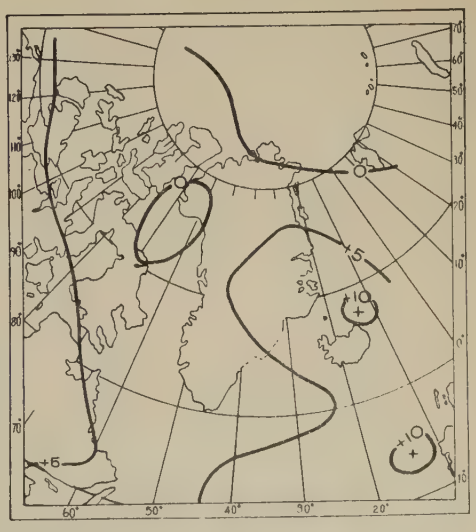

Figure 14. M.S.L. pressure anomaly, Jan. 1954.

(*b*) *Comparison with other years at other stations*

The precipitation totals for some Greenland stations are shown in Table 10.

TABLE 10. PRECIPITATION AT GREENLAND STATIONS

Station	Position	December 1952 (mm)	December 1953 (mm)	December Mean (mm)	January 1953 (mm)	January 1954 (mm)	January Mean (mm)	Years for Mean
Britannia Sø	77N 24W	Tr.	18·0		6·2	8·1		
Danmarkshavn	77N 19W	0·4	24	7	17	11	6	6
Thule	76N 67W	12	1	5	10	12	5	8
Myggbukta	73N 22W	0·1	44	45	37	47	56	11
Upernavik	73N 56W	20	4	10	14	15	12	5
Scoresbysund	70N 25W	19	96	65	24	31	25	7
Jakobshavn	69N 51W	22	7	13	15	8	11	4

It will be seen that the precipitation of December 1952 was very far below the normal and Dec. 1953 above the normal at east coast stations, but the January totals were not far from the normal. Conversely on the west coast precipitation was high in December 1952, low in December 1953 and normal for the January months.

It is concluded that the weather was finer than normal at Britannia Sø and probably also at Northice during December 1952 and that January 1953 was a fairly normal month.

ACKNOWLEDGMENTS

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551.510.52 : 551.515.82 : 551.557.5

Temperature, humidity and cloud near fronts in the middle and upper troposphere

By J. S. SAWYER
Meteorological Office, London

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Summary

Exploratory flights through fronts beneath the jet stream confirm that a narrow frontal zone some 30 to 50 mi wide with sharp boundaries is often present at the 500 mb level. On other occasions the temperature changes are more diffuse. The air in and near the frontal zone is usually very dry and must have undergone subsidence.

Dynamic considerations suggest that the very strong temperature gradient of the frontal zone at 500 mb arises from the differential subsidence of air on the cold side of the jet stream which leads to a tilting of the isentropic surfaces.

1. Introduction

Although from the initial introduction of the idea of a front into meteorology (Bjerknes 1918) the concept has been three-dimensional, nevertheless until recent years the weather forecaster has not been accustomed either to take into account the observed structure of fronts in the middle or upper troposphere, or even to relate the distribution of cloud at these levels to the position of the front in the upper troposphere. The brilliant analysis techniques of indirect aerology introduced by T. Bergeron (1928) are three-dimensional in concept, but still only relate phenomena of the surface weather map, and it is this chart which has been, and still is, the forecaster's main tool in deciding the distribution of cloud and rain.

On the other hand the present network of upper-air soundings enables reliable charts to be drawn for any selected level in the troposphere over much of the northern hemisphere, and attention has recently been directed to the well-defined nature of the frontal zone which can be identified upon these charts and, in particular, at the 500 mb level (Palmén 1951). Godson (1950) and others have gone further and suggested that an analysis of the three-dimensional shape of frontal surfaces should be made regularly by means of frontal contour charts and should form part of the routine at major forecast offices.

It therefore becomes desirable to study the character of the frontal zone in mid-troposphere, and the relation of the mid-troposphere front to the cloud and rain of frontal regions. It is also desirable to have some idea regarding the means by which the frontal zone is created and maintained. It was with these objects in view that a series of flights was planned by the Meteorological Research Flight of the Meteorological Office. Each flight was intended to cross the frontal zone at one or more levels in horizontal flight, and information on temperature, frost point and cloud were recorded. The flights were carried out in the years 1950 to 1955 and some of the results have already been published (Sawyer 1955).

In the present paper I wish to describe the observations so far available for the 500 mb and 600 mb levels, and to discuss their interpretation in terms of the dynamics of the jet stream and frontal zone.

2. The observations

A Hastings aircraft was used for the flights, and observations of temperature were made at half-minute intervals on a flat-plate electrical thermometer mounted beneath the nose of the aircraft. The error of the thermometer was probably about $\pm 0{\cdot}2°F$ and some doubt (of the order of 1°F) attaches to the correction for airspeed which should have been applied when the aircraft was in cloud. However, neither of these uncertainties can affect materially the temperature profiles shown in Figs. 1 and 2.

Frost point was obtained by means of a manually operated Dobson-Brewer frost point hygrometer. The probable error of these observations is less than 3°F, but the records are not so complete as those of temperature because of the difficulty of maintaining a long series of readings at only half-minute intervals. Visual observations of cloud were recorded on the flights.

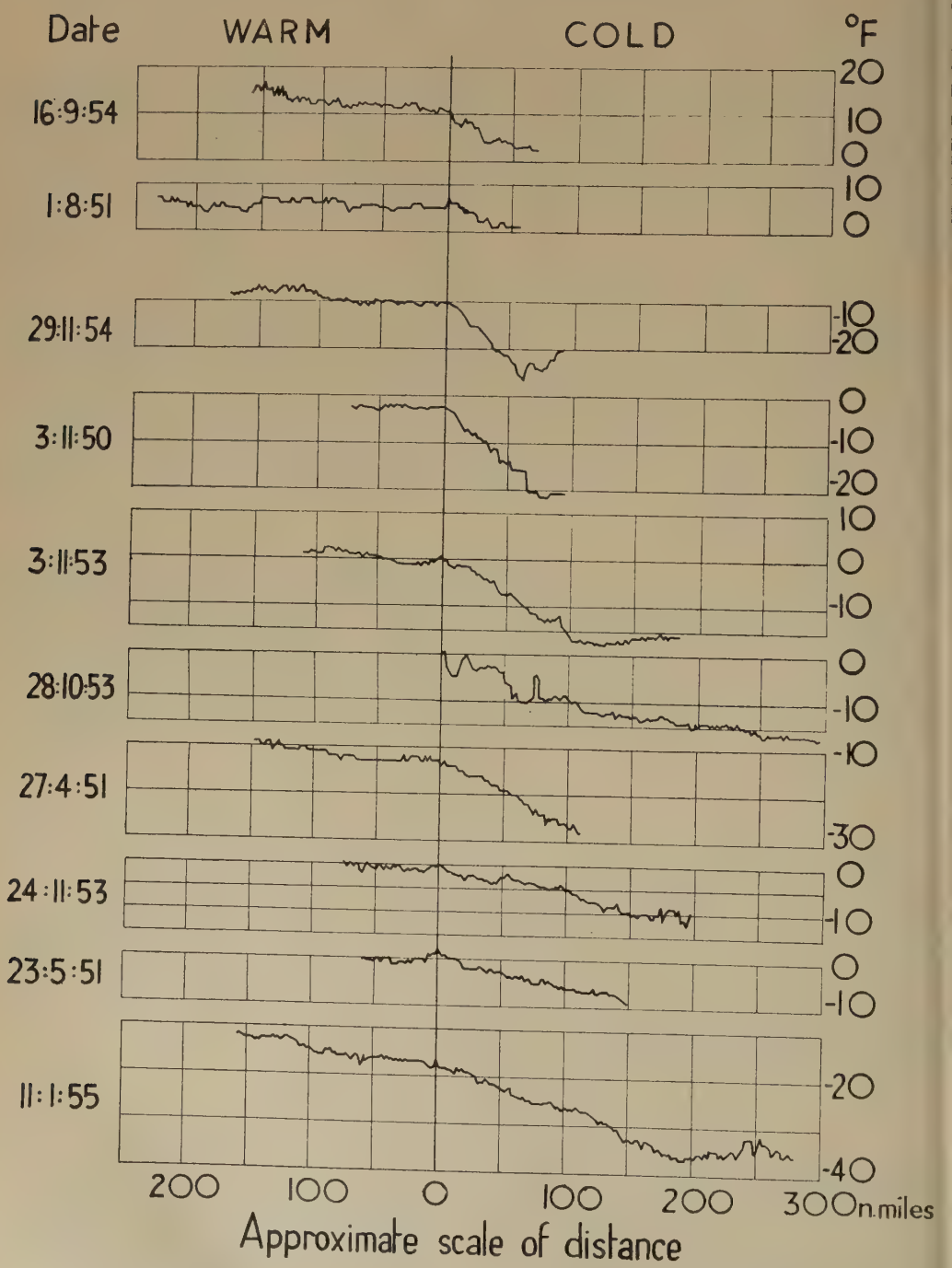

Figure 1. Horizontal distribution of temperature through frontal zones at 500 mb.

3. Temperature profiles

In order to examine the thermal structure of fronts in mid-troposphere the temperature observations made during all horizontal traverses of fronts at or near the 500 mb and 600 mb levels have been collected in Figs. 1 and 2. In these diagrams the temperature has been plotted against time but since the aircraft speed was usually around 220 kt and

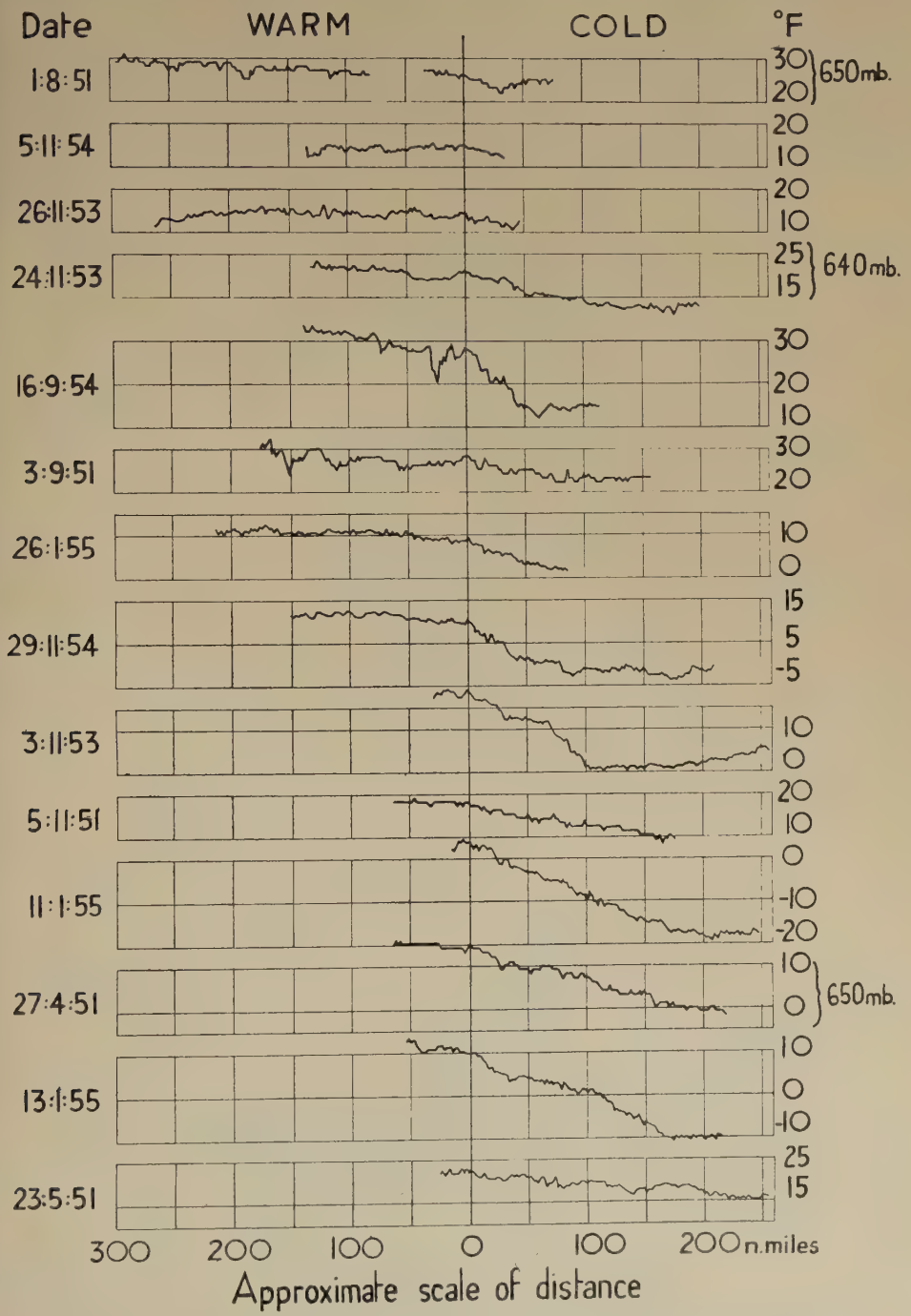

Figure 2. Horizontal distribution of temperature through frontal zones at 600 mb.

the flight track roughly perpendicular to the frontal zone, the plotted temperatures ma be regarded as displaying the horizontal profile of temperature across the front with th scale of distance indicated at the foot of the diagram. One point of the temperatur profile has been identified as corresponding to the warm boundary of the frontal zone and the temperature plots have been aligned for convenience with these points in a vertica line. All flights which intersected an identifiable front in mid-troposphere on a cross section based on radiosonde observations have been included. Data for some of th flights have already been published (Sawyer 1955); other observations come from flight not yet fully analysed. On some occasions level runs were made near both 500 mb an 600 mb levels, but on other occasions only one level in mid-troposphere was included

At 500 mb, a well-defined frontal zone with quite sharp changes of temperatur gradient can be identified on about half of the occasions. On the remaining flights th

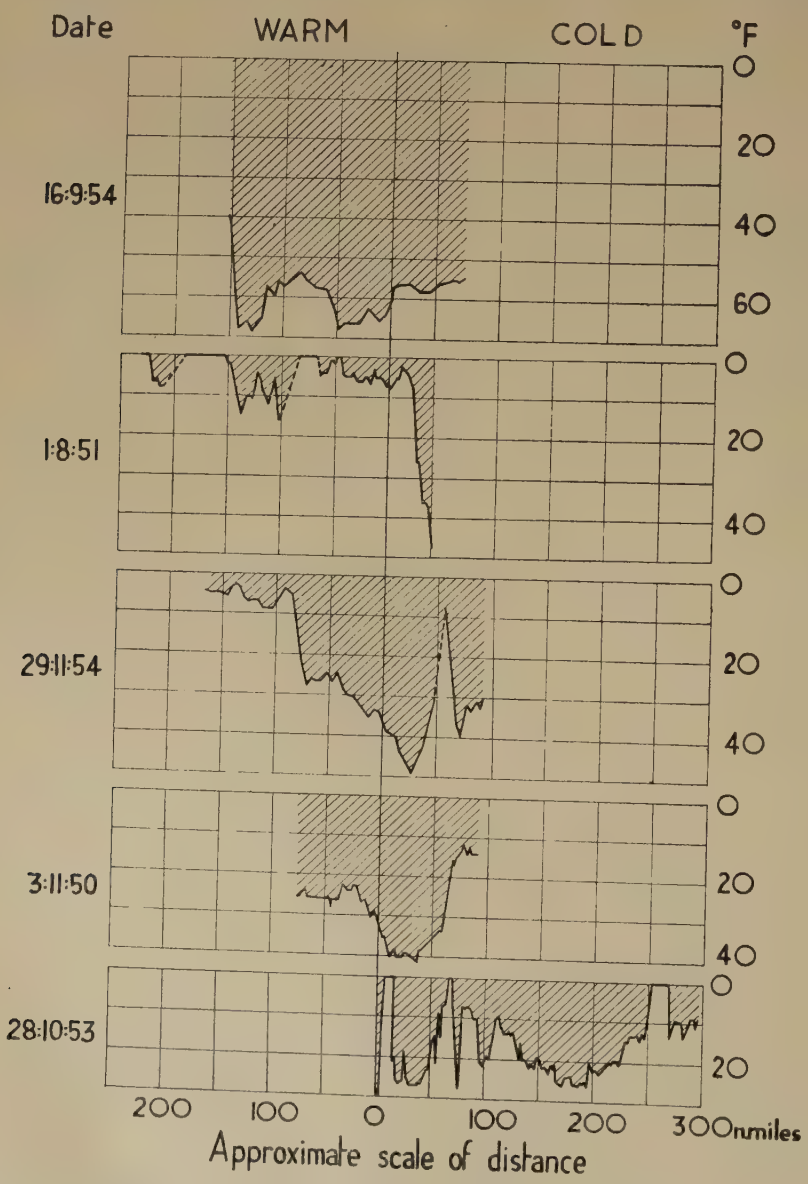

Figure 3 (a). Horizontal distribution of frost-point depression through frontal zones at 500 mb.

egion of temperature change is more or less diffuse. The width of the frontal zone as bserved on these flights varied between 35 and 170 naut. mi, and the temperature change rom about 6°F to 18°F. The strongest maintained temperature gradient was that on 9 November 1954 which amounted to 10°F in 35 naut. mi.

At 600 mb the temperature profiles are similar but, on the whole, rather less clearly lefined. On less than a third of the temperature profiles in Fig. 2 can the frontal zone be dentified unequivocably. This may be purely an accident of the sample but, more pro-ably, there is a greater tendency to form a narrow frontal zone with clear-cut boundaries t 500 mb.

4. Distribution of humidity

Figs. 3 and 4 show the horizontal variation of frost-point depression at the 500 mb nd 600 mb level. With one exception for which no hygrometer observations are available

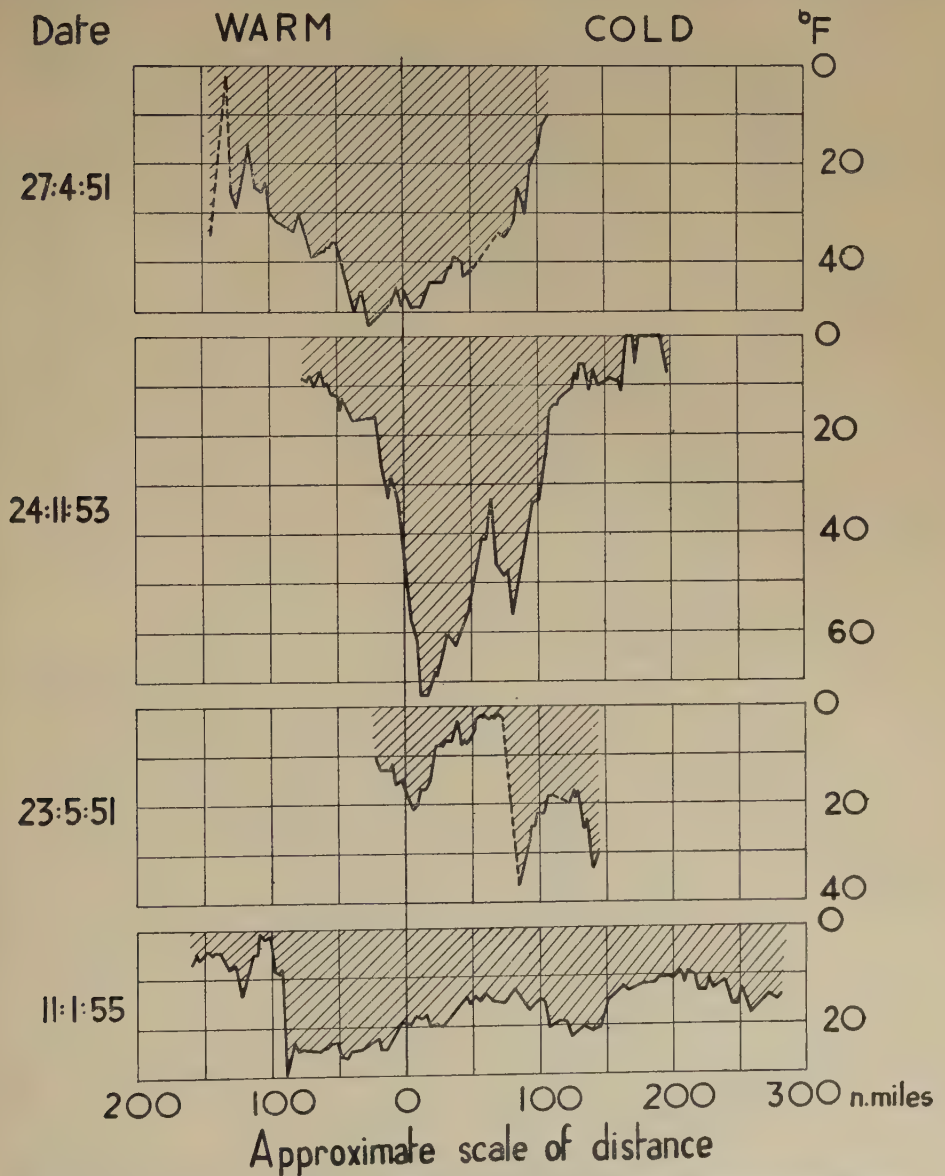

Figure 3 (b). Horizontal distribution of frost-point depression through frontal zones at 500 mb.

these correspond to the temperature observations in Figs. 1 and 2. The frost-point plots are aligned with respect to the warm boundary of the frontal zone as defined by the temperature observations.

The most striking feature of these humidity observations is the very dry air which is frequently, but not always, found in and near the frontal zone. On seven out of the nine available frost-point traverses at 500 mb, the frost-point depression exceeds 30°F and on two occasions frost-point depressions exceeding 60°F were measured. On the flights which failed to encounter this very dry air, the air is still not really moist. Frost-point depressions less than 10°F seem to be the exception and cloud in the frontal zone was encountered only once. On two other occasions the aircraft flew through cloud some distance on the warm side of the frontal zone, and once on the cold side. Cloud would no doubt have been encountered more frequently if the flights could have been extended for greater horizontal distances into the warm air.

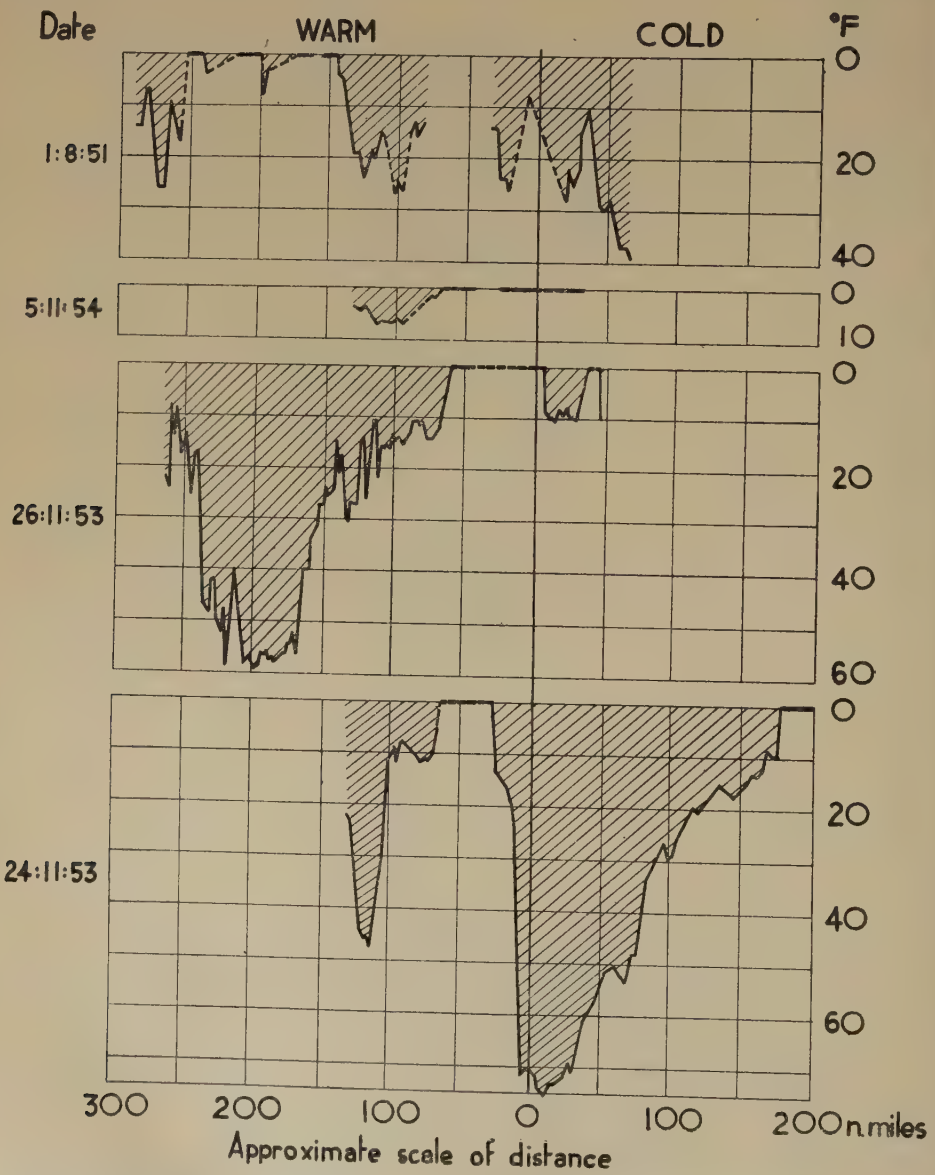

Figure 4 (a). Horizontal distribution of frost-point depression through frontal zones at 600 mb.

The distribution of the frost-point depression does not appear to be closely related to the frontal zone or its boundaries, except that when the frontal zone was very clearly defined by the temperature profile (e.g., 29 November 1954, 3 November 1950, and 27 April 1951) the very dry air was present and the frost-point depression reached its extreme near the warm boundary of the frontal zone.

At 600 mb the humidity distribution seems even more erratic than at 500 mb. Some very dry air (frost-point depression greater than 30°F) occurred on 7 out of the 12 flights and on most of them frost-point depression is greater than 10°F over considerable stretches of the flight. On three of the flights the very dry air was closely associated with the frontal zone but on the remainder it was irregularly distributed.

The aircraft passed through cloud on 9 out of the 12 flights. On six of these the cloud did not significantly enter the frontal zone, being mainly on the warm side. On

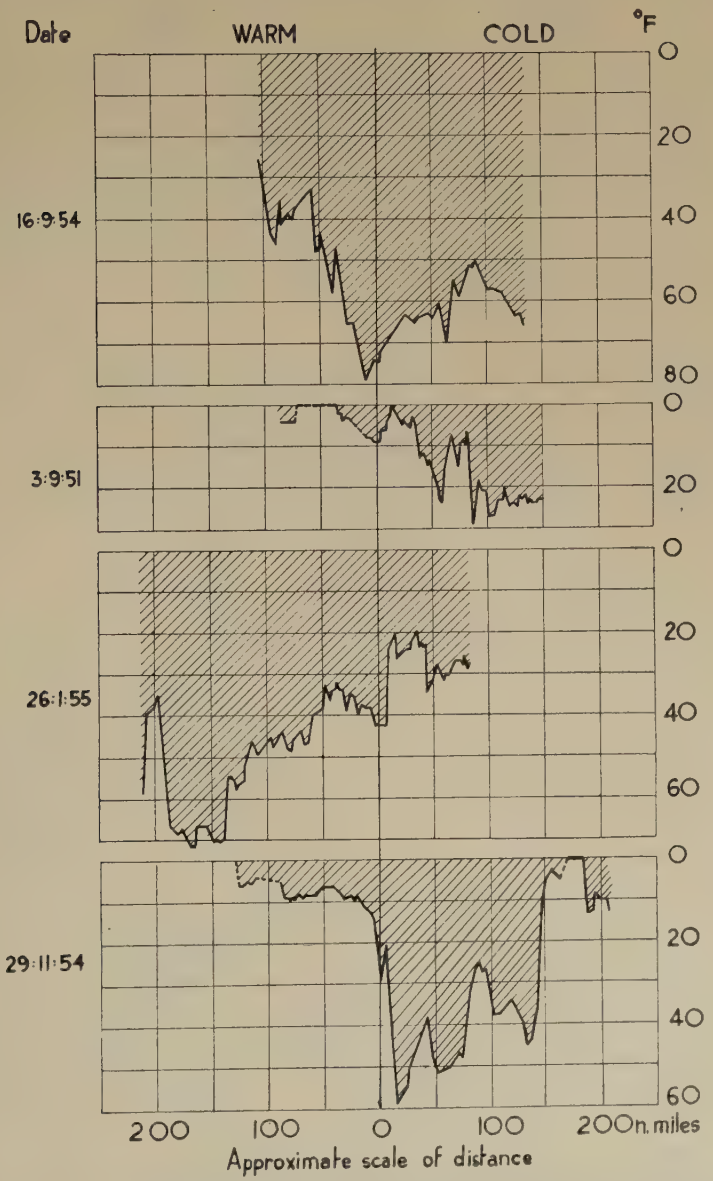

Figure 4 (b). Horizontal distribution of frost-point depression through frontal zones at 600 mb.

three flights, however, there was extensive cloud in the frontal zone and on one of thes solid cloud appeared to fill most of the transitional region. This occurred in a broa warm-front zone with a steep slope of 1 in 30 and might be regarded as exceptional.

5. Summary of the observations

In order to summarize the data, average temperature and frost-point profiles have bee calculated (Figs. 5 and 6). This has been done using the position of the warm boundar of the frontal zone as a reference point, and the temperature at this point as a referenc temperature. In view of the wide dissimilarity of individual profiles the resulting averag cannot be regarded as particularly representative. However, it serves to emphasize tha at 500 mb the air in and surrounding the frontal zone is normally dry and has the lowes

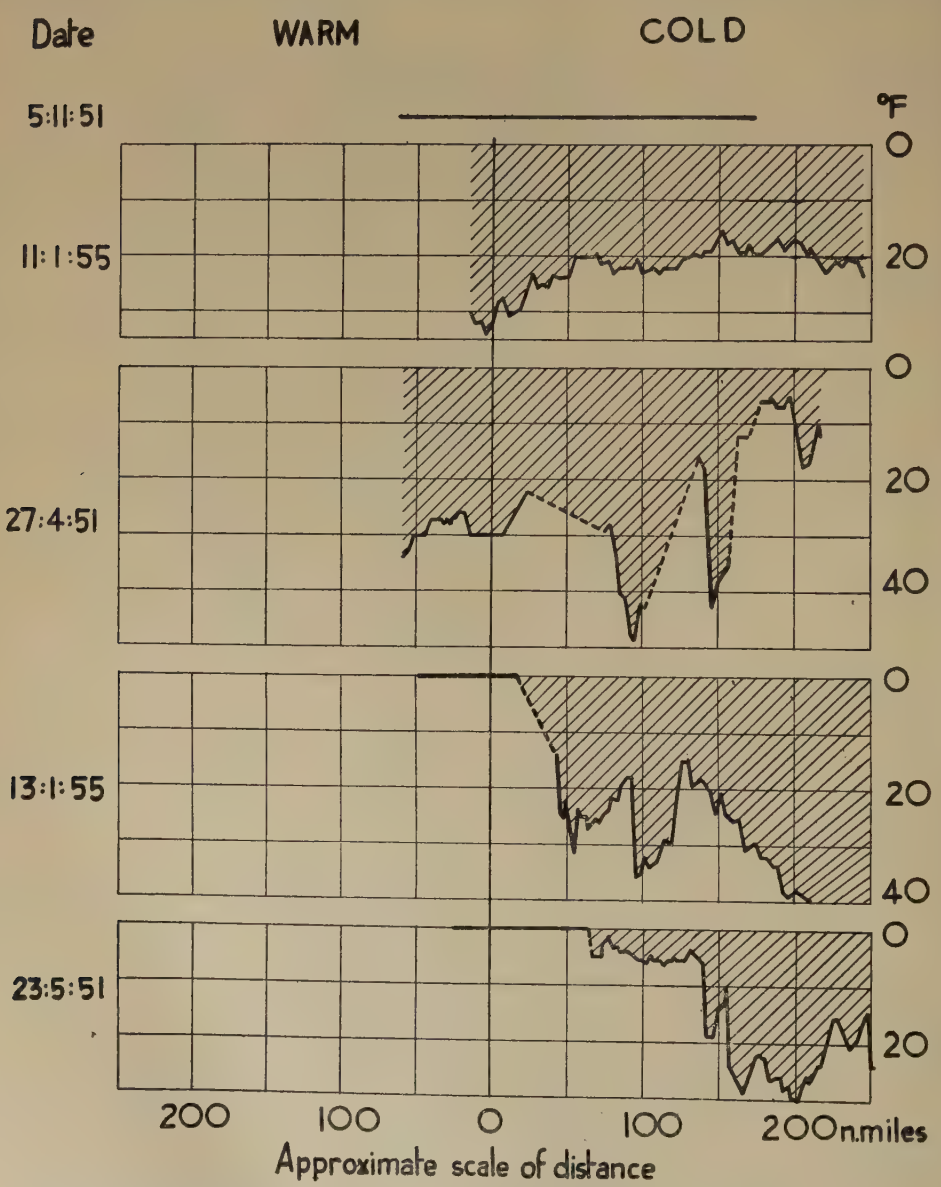

Figure 4 (c). Horizontal distribution of frost-point depression through frontal zones at 600 mb.

relative humidity near the warm boundary of the frontal zone. The average frost-point depression everywhere exceeds 15°F and reaches 35°F in the frontal zone. There is no sign of a rapid increase in water content across the frontal zone or its boundaries and it seems quite clear that at this level the frontal zone does not represent a transition from a cold dry air mass to a warm moist one; nor is the idea of up-gliding of air immediately above the frontal zone substantiated.

At 600 mb the average structure is somewhat different. The average frost-point depression is relatively small just on the warm side of the frontal zone (less than 10°F), while a dry region occurs in the frontal zone (in which the frost-point depression exceeds 20°F). The moist region just in the ' warm air ' adjacent to the front is found on individual flights and saturation was reached in this region on several occasions with accompanying cloud.

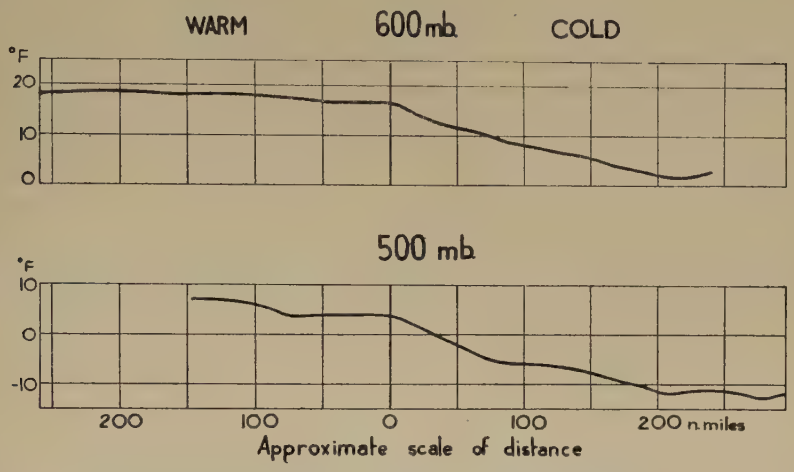

Figure 5. Average horizontal temperature profile across frontal zones at 500 mb and 600 mb.

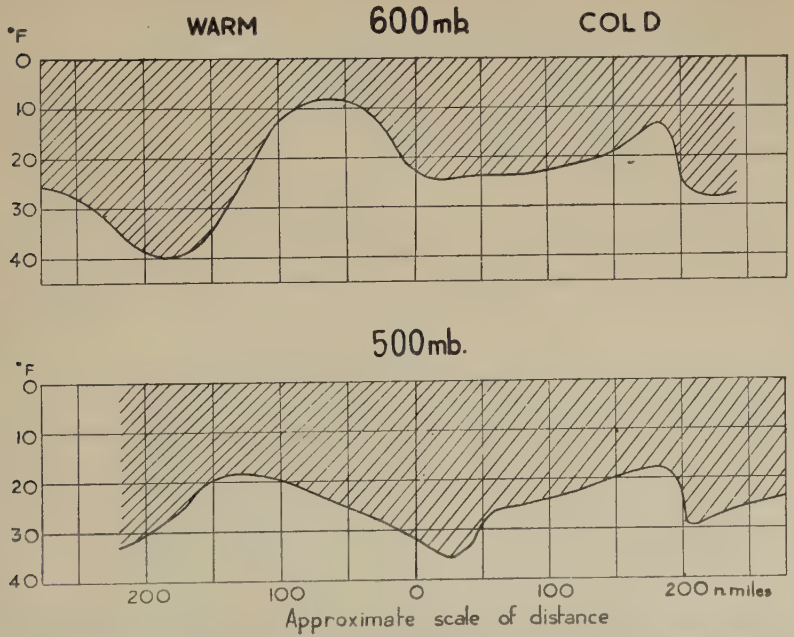

Figure 6. Average horizontal profile of frost-point depression across frontal zones at 500 mb and 600 mb.

6. Some theoretical considerations

Fronts are closely linked with depressions and with the jet stream, and any completely satisfying explanation of the character of fronts would probably be possible only as part of a more comprehensive theory of depressions and the jet stream. Nevertheless there is value in more limited theoretical discussions, and in the following paragraphs I have attempted to show that some of the characteristics of fronts in the upper troposphere follow as consequences if we accept the existence of the jet stream, and the fact that it has a coherent and only slowly varying structure. Thus I wish to side-step the question 'Why is the front there?' and merely consider why some of the properties which we observe are necessary for its self-consistency.

For this purpose it is useful to consider the conditions under which a jet stream can persist in the upper troposphere in only slowly varying form. It is a fact of observation that this occurs, and that the jet-stream current is sinuous and that the velocity in the core varies along its length. The troughs and ridges and the velocity maxima move considerably more slowly than the wind. For simplicity, we may regard the flow pattern as stationary with respect to the flow at some level in mid-troposphere. To be specific we regard the flow as steady relative to co-ordinates moving with the 600 mb flow, and we examine the changes of vorticity which occur at higher levels as the air flows through the pattern.

Cross-sections drawn through the jet stream draw attention to the narrow character of the velocity maximum and the high values of horizontal wind shear, more particularly on the cold side. A typical wind profile would be that represented in Fig. 7 (a) and the corresponding vorticity profile based on the wind shear would be as represented in Fig. 7 (b).

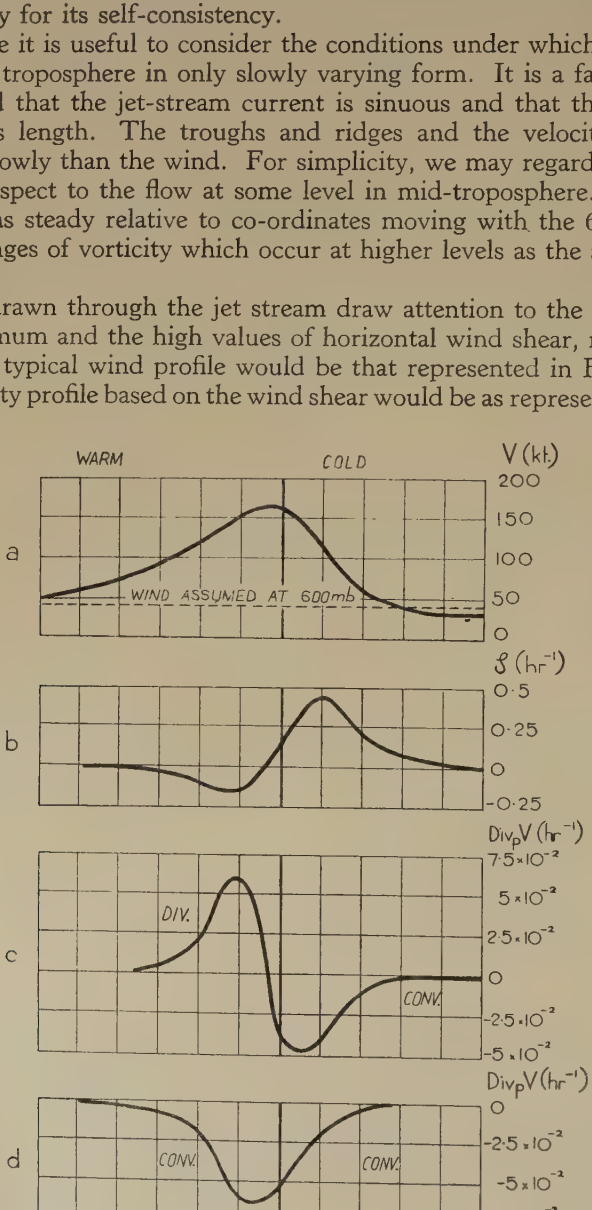

Figure 7. (a) Schematic profile of wind through a jet stream; (b) corresponding profile of relative vorticity; (c) distribution of divergence required if the jet stream intensifies along its length; (d) distribution of divergence required if the jet stream is to acquire a curvature with 600 mi. radius in a distance of 1,000 mi.

Changes in the intensity of the jet stream or changes of its curvature along its length will involve changes in the vertical component of vorticity, and it is useful to examine how these arise. The procedure which I shall adopt is to assume (1) that the jet stream intensifies without change in the shape of wind profile (that is the wind increases in the same proportion at all points across the stream) or, alternatively, (2) that from being straight the jet becomes curved downstream (again without change of form), and I shall examine the pattern of divergence and vertical motion which would be necessary to satisfy the vorticity equation. Regarded strictly this procedure is invalid because the wind profile may, and probably does, change along the stream but the only alternative appears to be to attempt the solution of the three-dimensional dynamical problem. The simpler approach is both more practical, and shows more clearly the interaction of the various dynamical processes.

The vorticity equation in isobaric co-ordinates can be used to describe the change of vertical component of vorticity as the air flows along the jet stream. Thus

$$\frac{D\zeta}{Dt} - \frac{\partial\omega}{\partial y}\frac{\partial u}{\partial p} + \frac{\partial\omega}{\partial x}\frac{\partial v}{\partial p} + (l + \zeta)\,\mathrm{Div}_p\,\mathbf{v} = 0 \qquad . \qquad . \qquad (1)$$

where $\omega \equiv Dp/Dt$

If we choose the x-axis along the direction of the jet stream, $\partial u/\partial p$ is very much greater than $\partial v/\partial p$, and we can neglect the term $\dfrac{\partial\omega}{\partial x}\dfrac{\partial v}{\partial p}$.

If, further, we assume that the flow is steady relative to some co-ordinate system moving with the general features of the flow at a speed corresponding to the wind at say 600 mb, and we ignore the local vorticity change in this moving co-ordinate system compared with the advective change, then

$$V\frac{\partial\zeta}{\partial s} + \omega\frac{\partial\zeta}{\partial p} - \frac{\partial\omega}{\partial y}\frac{\partial u}{\partial p} + (l + \zeta)\,\mathrm{Div}_p\,\mathbf{v} = 0 \qquad . \qquad . \qquad (2)$$

where V is the wind velocity relative to the wind at 600 mb and $\partial\zeta/\partial s$ is the rate of vorticity change along a horizontal projection of a stream line.

In order to make some general deductions about the character of the flow we will assume that $V\,\partial\zeta/\partial s$ can be roughly estimated from the fact that the jet is assumed (1) to intensify downstream or, alternatively, (2) to acquire curvature along its length. We may also assume from the presence of relatively undistorted isentropic surfaces in the stratosphere only a little above the jet that there is a level of small vertical motion somewhere near the jet level. At such a level the principal terms of Eq. (2) must be $V\,\partial\zeta/\partial s$ and $(l + \zeta)\,\mathrm{Div}_p\,\mathbf{v}$. The divergence, $\mathrm{Div}_p\,\mathbf{v}$, can be calculated from these two terms alone.

On this basis, the distribution of $\mathrm{Div}_p\,\mathbf{v}$ across the jet stream must be broadly as indicated in Figs. 7 (c) and 7 (d) for the two assumptions of a jet stream (1) intensifying along its axis and (2) acquiring cyclonic curvature downstream. The particular conditions assumed in deriving Figs. 7 (c) and 7 (d) respectively are (1) that the jet-stream velocities increase by 10 per cent in 100 mi downstream $\left(V\dfrac{\partial\zeta}{\partial s} = V\zeta \times 10^{-3}\ \mathrm{hr}^{-2}, \text{ with } V \text{ in kt}\right)$ and (2) that the jet stream acquires a curvature with 600 naut. mi radius at a distance 1000 naut. mi downstream from a straight section.

An interesting feature of these distributions of divergence is the strong gradients of divergence across the jet stream, and in particular the strong increase of convergence towards the jet axis from the cold side. If the vertical velocity is small near the jet-stream level, then at lower levels towards mid-troposphere a strong gradient of vertical velocity must be expected with strong descent near to the jet-stream axis and relatively weak descent only some 200 mi further towards the cold side of the jet. This differential subsidence can be expected to be effective in setting up a strong horizontal gradient of temperature in mid-troposphere as observed in the 500 mb frontal zone.

However, the picture is more complicated because the resulting values of the vertical velocity are large enough to make the second and third terms in Eq. (2) comparable in magnitude with the others. (The second and third terms represent respectively the vertical advection of vorticity and the vorticity change arising from the tilting of the vortex tubes). Were in fact the values of the divergence represented in Figs. 7 (c) and 7 (d) continued down to the 500 mb level then there would be values of ω and $\partial\omega/\partial x$ occurring at that level of as much as 10 mb hr^{-1} and $3{\cdot}5 \times 10^{-2}$ mb hr^{-1} naut. mi^{-1}. If we assume that the wind increases linearly from a uniform value of 40 kt at the 600 mb level up to the level of the jet stream at say 300 mb, then the order of magnitude of the terms in Eq. (2) would be as indicated in Table 1. The crude assumption has been made that values of ω can be calculated from the divergence obtained ignoring the terms $\omega\,\partial\zeta/\partial p$ and $-\frac{\partial\omega}{\partial y}\frac{\partial u}{\partial p}$.

TABLE 1. RELATIVE MAGNITUDE OF TERMS IN EQ. (2)

		$V\frac{\partial\zeta}{\partial s}$	$\omega\frac{\partial\zeta}{\partial p}$	$-\frac{\partial\omega}{\partial y}\frac{\partial u}{\partial p}$	$(l+\zeta)\,\mathrm{Div}_p\,\mathbf{v}$
		(Units hr^{-2})			
400 mb	50 mi from axis	$1{\cdot}6 \times 10^{-2}$	$-0{\cdot}4 \times 10^{-2}$	0	$-1{\cdot}2 \times 10^{-2}$
	150 mi from axis	$0{\cdot}7 \times 10^{-2}$	$-0{\cdot}2 \times 10^{-2}$	$-0{\cdot}2 \times 10^{-2}$	$-0{\cdot}3 \times 10^{-2}$
500 mb	50 mi from axis	$0{\cdot}4 \times 10^{-2}$	$-0{\cdot}8 \times 10^{-2}$	0	$+0{\cdot}4 \times 10^{-2}$
	150 mi from axis	$0{\cdot}2 \times 10^{-2}$	$-0{\cdot}4 \times 10^{-2}$	$-0{\cdot}4 \times 10^{-2}$	$+0{\cdot}6 \times 10^{-2}$

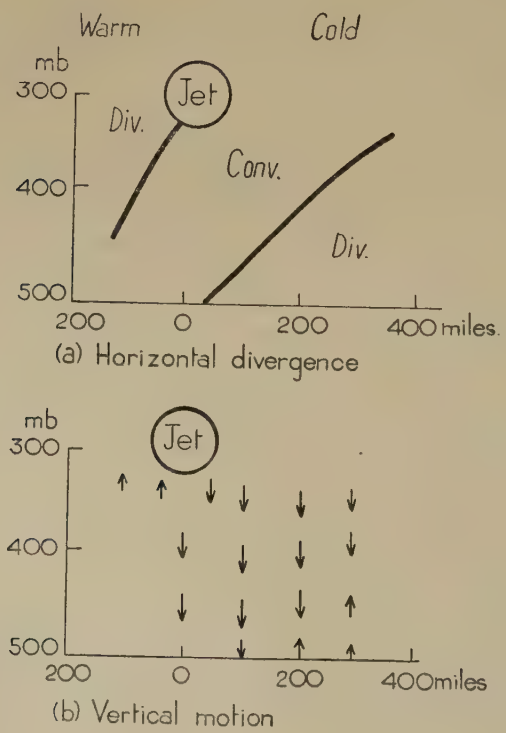

Figure 8. Distribution of horizontal divergence and vertical motion on the cold side of an intensifying jet stream.

It is immediately clear that both the vertical advection of vorticity, $\omega\, \partial\zeta/\partial p$, and the 'tilting term,' $-\partial\omega/\partial y\, \partial u/\partial p$, are likely to be significant in the region immediately below and on the cold side of the jet stream. This has been confirmed, particularly as regards the latter term, by individual case studies by Reed and Sanders (1953), Reed (1955) and Newton (1954).

The values of ω on which the preceding figures are based assume that the convergence can be calculated ignoring the two terms which we have seen to be important. This is clearly invalid. In fact we can deduce that the convergence rapidly decreases downwards, and may even be replaced by divergence. It is also clear that this reversal should take place more rapidly at 150 mi from the jet axis than closer to the axis because the stronger horizontal gradient of ω at the former position gives rise to larger values of the 'tilting term.' The fields of divergence and of vertical motion in the region below and on the cold side of the jet must be roughly as indicated in Figs. 8 (a) and 8 (b).

The region of horizontal divergence below and to the cold side of the jet stream will bring about an intensification of the vertical gradient of potential temperature (i.e., it will increase the vertical stability) in this region as is observed in the 500 mb frontal zone. At the same time the tilting about a horizontal axis will increase the horizontal temperature gradient, and this will be accompanied by an increase in the cyclonic vorticity about a vertical axis apparently primarily as a result of the 'tilting term,' $-\partial\omega/\partial y\, \partial u/\partial p$ in Eq. (1).

It is not profitable to pursue the argument further without a more satisfactory treatment of the balance between the various dynamical factors involved. Nevertheless, the treatment is probably adequate to show that a jet stream which has a strong shear on its cold side, and flows from a region of relatively low speed to one of higher speed, must develop a region of general subsidence on its cold side which is accompanied by an intensification of both horizontal and vertical gradients of potential temperature such as is observed in 500 mb frontal zones.

Similar conclusions would follow regarding regions where the jet stream flows into a slow-moving upper trough although the frontal zone would then be built up at a smaller horizontal distance from the jet axis.

This dynamical process provides no explanation as to why the boundaries of frontal zones should be so sharp and well-defined as is often observed. However, these rather abrupt changes of temperature gradient may well arise from the intensification and tilting of already existing boundaries to inversions or stable layers. The horizontal divergence will intensify the changes in vertical gradients of temperature and the tilting will introduce into the horizontal temperature gradients much stronger discontinuities than are normally present. Reed and Sanders (1953) describe a case in which the subsidence was sufficiently great to bring stratospheric air into the frontal zone, and the tropopause itself provided the initial discontinuity of lapse rate which became the boundaries of the frontal zone. I also examined a similar case which occurred over north France on 3 November 1950 (Sawyer 1955) and which appears in Figs. 1 and 3. However, these cases are probably exceptional, and stable layers and inversions in the troposphere may also provide the initial discontinuities of lapse rate which are intensified into the boundaries of frontal zones.

So far, I have discussed the formation of a frontal zone in intense sections of the jet stream and in the regions of upper troughs. The arguments may be reversed to suggest a reverse circulation and frontolysis as the air passes into weaker sections of the jet and into upper ridges. No doubt detailed observations of the jet stream and studies of upper fronts have concentrated on regions where they are intense, but I think there is another reason why the structure which is built up in these intense regions of the jet and frontal zone should persist elsewhere. That is because it is unlikely that the vertical circulation as the air enters a diffluence or a ridge will be the exact reverse of that in a preceding confluence or trough. It will not be the same air which descended in the confluence that ascends again in the diffluence – sufficient small changes in the relative position of front and jet will probably have taken place to mean that the air which rises most will not be

precisely that which descended most. Thus in association with weaker sections of the jet and with upper ridges, we might expect to find some remnant, albeit less intense, of the dry frontal zone built up in a preceding confluence or trough.

7. Conclusions

The temperature observations on horizontal traverses beneath the jet stream confirm that a narrow frontal zone, some 30 to 50 mi wide, with sharp boundaries is often present at the 500 mb level. On other occasions the temperature changes are more diffuse. Humidity observations indicate that the air in and near the frontal zone has undergone subsidence and make it seem unlikely that the frontal zone at this level owes its origin to the different geographical origin of the air on either side of it. Theoretical considerations suggest the horizontal temperature gradient arises from the differential subsidence of air on the cold side of the jet stream which leads to tilting of the isentropic surfaces. Dynamical considerations also explain the vertical stability and cyclonic vorticity associated with the front.

The front-forming processes in the upper troposphere only operate where the jet stream is intensifying or becoming more cyclonically curved along its length, but the characteristic structure is carried forward into other weaker sections of the jet and probably does not degenerate completely. This may account for the variety of temperature and humidity profiles observed.

At 600 mb the temperature gradients at fronts are somewhat less well-defined than at 500 mb, and although subsided air is often found in the frontal zone, the air on the warm side of the warm frontal boundary is usually relatively moist. Saturation and cloud occur in this region and the structure corresponds more closely at 600 mb to the conventional front with upsliding warm air. This conforms with previous findings (Sawyer and Dinsdale 1955) that the boundary of the frontal cloud mass slopes more steeply than the front itself and on the average intersects the warm boundary of the frontal zone around 600 mb.

Acknowledgment

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551.557.5 : 551.513.1 : 608.4

The general circulation in a steady rotating-dishpan experiment

By HERBERT RIEHL and DAVE FULTZ
University of Chicago

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SUMMARY

This is the continuation of a previous paper (*Q.J.*, **83**, 1957) in which the structure of the steady rotating-dishpan experiment was described. The present article begins with a calculation of the heat balance and a determination of the mechanisms that transport heat from heat to cold source. Of particular interest is the question whether the heat transport is effected by quasi-geostrophic horizontal or by non-geostrophic vertical circulations. Calculations were performed in the conventional polar coordinate system and also in a curvilinear coordinate system following the jet-stream axis. In polar coordinates the heat transport is accomplished largely by horizontal circulations, and in the curvilinear coordinate system by an ageostrophic vertical mass circulation. Thus the ' mechanism ' of heat transfer that is found depends largely on the framework of reference in which calculations are made.

When the vertical circulations are averaged with respect to longitude in the two coordinate systems, a three-cell structure with direct equatorial and polar branches and an indirect middle-latitude branch is found in polar coordinates. In contrast, the calculations yield a single direct cell for the curvilinear system.

Computation of the transport of absolute angular momentum in the two coordinate systems produces corresponding results.

Calculations of kinetic and potential energy balance were performed in curvilinear coordinates. Production of kinetic energy by the mass circulation far exceeds actual net generation of kinetic energy, so that ' eddies ' with respect to the jet-stream axis convert kinetic energy to work done across the boundaries of the volume considered. Similarly, the potential energy release is achieved by means of the mass circulation while the eddies store potential energy poleward of the jet-stream axis.

An attempt was made to determine whether the velocity distribution near the top surface, especially the jet-stream core, could be derived using only the mass circulation and the theorem of conservation of potential vorticity. The result showed that the velocity profile on the anticyclonic side of the jet stream is well reproduced by the calculation, but that the computed jet-stream axis is situated far poleward of the observed position and that the computed strength of the jet centre exceeds the actual strength by 50 per cent.

In conclusion, the result of the computations shows that at least two, and possibly a much greater number of general circulation ' mechanisms ' can be found, depending upon the reference frame chosen. Mathematically, these solutions are all equally correct and yield the same total transports. From the physical viewpoint, the general circulation is represented as a simple heat engine in the curvilinear coordinate system. Considering the energy balance, the solution indicates that the kinetic energy of the general circulation is produced and maintained by the largest cell, the single direct circulation extending from heat to cold source, and that this kinetic energy is then converted to kinetic energy of smaller circulations.

1. INTRODUCTION

In a previous article (Riehl and Fultz 1957, hereafter referred to as I) the structure of the circulation in a steady rotating-dishpan experiment with heat and cold source was described. The principal feature of the experiment is a well-marked jet stream which forms three progressive long waves. This jet stream has about the same strength relative to the equatorial speed of the pan as atmospheric jet streams of the high troposphere have relative to the equatorial speed of the earth. The temperature decreases from heat to cold source at all levels, and isotherms are in phase with streamlines in the upper part of the fluid where the jet stream is located. This current, which is nearly geostrophic, thus decreases in intensity downward. In the lower part of the fluid we observe instead a cyclone and an anticyclone, situated respectively under the inflection points of the jet axis (southwest and northwest winds), in accord with atmospheric structure.

The dynamic principle guiding the motion is conservation of potential vorticity. In the jet-stream core, divergence and vertical motion are small; here the principle of conservation of absolute vorticity applies approximately. Under the assumption of adiabatic (isothermal) motion in the interior of the fluid, vertical motions were calculated.

As in the atmosphere, ascent occurs east of the low-level cyclone while descent pre dominates in the region of strong equatorward flow between cyclone and anticyclone After averaging the vertical motion around latitude circles, net ascent is found poleward and net descent equatorward of the strongest westerlies. However, when averaging i performed following the jet axis, exactly the inverse result is obtained. This suggest strongly that the methods employed in summarizing features of the flow into a genera circulation picture can have a profound effect on the result; hence much care is necessary in assessing the scope of conclusions that may be drawn from any particular set of compu tations.

The main objective of this report is to inquire into the mechanisms maintaining the general circulation described in (I). We shall consider the transports of heat and absolute angular momentum. This will be followed by computations of the release of kinetic and potential energy, and by some remarks about the jet stream and the energy cycle of the general circulation.

2. Heat sources, sinks and transfer mechanisms

(a) *Heat sources and sinks*

As stated in (I) the pan has a radius of 15·6 cm at the water surface and the water depth is 4 cm. A heat source with nominal intensity of 150 w, most of which went directly to the laboratory atmosphere, was applied at the equatorial rim, and a cylindrical central cold source with radius of 6·5 cm provided for balance. This cold source was operated by means of circulating cold water with temperature of about 10°C, while the mean temperature of the water in the pan was about 19·5°C. The heat transported by the working liquid from heat to cold source was computed by measuring the temperature of the water

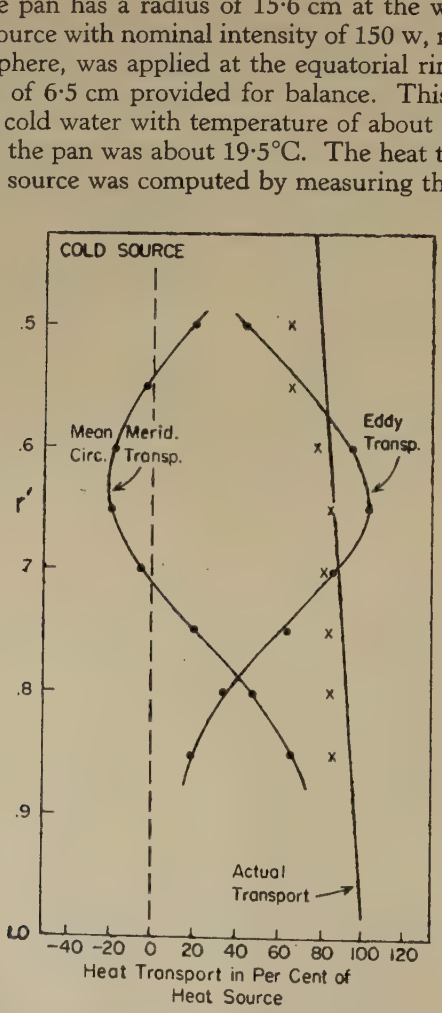

Figure 1. Heat transport in cylindrical coordinates, in per cent of the equatorial heat source. Straight line indicates actual heat flow, crosses indicate computed heat flow.

before entering and after leaving the cold source. Certain extraneous gains of heat at the cold source arose from conduction in the metal at the bottom, contact of the cold source with the atmosphere, and from the lights needed for photography. These gains were estimated from heat flows measured at the cold source in several null experiments without water in the pan and gave a corrected abstraction of heat by the cold source from the water in the pan of 45 w.

On account of the laboratory conditions prevailing at the time of the experiment, evaporation from the pan into the atmosphere did occur. Total evaporation was computed from the volume change of the water through the experiment; the result indicates a cold source of 9 w. The heat source therefore had an effective strength of 54 w. Since the radial distribution of evaporation is not known, it has been assumed to be linear, an unimportant assumption as will be seen from Fig. 1.

In the experiment the equatorial rim was not insulated so that heating spread along the bottom of the pan. Thus there are two convective layers of which one, near the bottom, is produced by heating and the other, near the top, by evaporation cooling. Bottom heat source and top cold source did not have much influence on the structure of the general circulation. The experiment has been repeated in a pan with insulated bottom and with a glass top shielding the water from the atmosphere. In this way both heat loss at the top and heat gain at the bottom were much reduced. Only minor modifications occurred in the new experiment compared to the previous one which forms the basis for this study. It therefore will be assumed in the following that heat and cold source can be considered as being applied only at equatorial and polar rims of the basin. Validity of this assumption is also supported by the fact that strong stability prevailed throughout the bulk of the water and that the temperature profile was nearly linear with height except near top and bottom (I). For these reasons it is unlikely that appreciable vertical heat diffusion occurred, at least over the latitude range for which computations were made.

(*b*) *Heat-transfer mechanisms*

We shall consider a coordinate system fixed in the waves and travelling relative to the pan with the speed of the waves. The subscript r on velocity components will refer, as in (I), to motion relative to this wave coordinate system as distinct from motion relative to the pan.* The convective heat transfer (H) across an arbitrary closed wall is given by

$$H = \oint_s \int_0^D (c_w T)(\rho c_n)\, ds\, dz \qquad . \qquad . \qquad . \qquad (1)$$

Here s is the horizontal coordinate following the boundary, z the vertical coordinate, c_n the horizontal velocity component normal to the wall considered positive when directed from heat to cold source (inward), T temperature, c_w specific heat of water, ρ density and D the depth of the fluid. Density and specific heat may be taken as unity in evaluating Eq. 1. The conductive contributions to the total heat transfer are negligible, being of order 1/60 of the convective contribution. The question of main importance, following recent literature (Bjerknes 1951, Starr 1948, Rossby 1949), is whether the transfer of heat, as also that of momentum and other properties, is carried out mostly by ageostrophic or quasi-geostrophic processes. Non-geostrophic transfer is produced by a circulation in the vertical plane about a horizontal axis parallelling the boundary of integration in Eq. 1 where we now specify the boundary to be a vertical wall whose trace is one of the curvilinear coordinate lines of Fig. 8. It depends on the existence, at least at some levels, of a

* Three coordinate systems are used in I and in the present paper. The first is the usual cylindrical coordinate system rotating at the same rate as the container. The second is a 'wave' coordinate system consisting of cylindrical coordinates whose absolute rate of rotation is the same as that of the waves. This system is identified by subscript r alone, standing for 'relative to waves.' The last system is a curvilinear one (s, n, z) rotating at the same rate as the waves. The tangential s and normal n directions are respectively parallel to the jet-stream axis at the top and orthogonal with positive direction toward the cold source (see Fig. 8). Subscripts rs or rn identify components with respect to this frame of 'jet coordinates.'

net mass flow $\oint c_{rn}\, ds = \bar{c}_{rn} \cdot s$, where $\bar{c}_{rn} = \bar{c}_{rna}$, the average ageostrophic mass flow across the boundary. Because of the mass continuity requirement $\oint_s \int_0^D c_{rn}\, ds\, dz = 0$, so that for net mass flow in any layer a compensating opposite mass flow must occur in another layer. If mass flow toward the cold source occurs near the top and toward the heat source near the bottom, the mass circulation will transport heat from heat source to cold source because the stratification of the water is stable, i.e., the temperature of the water, integrated around the boundary, is higher at the top than at the bottom.

The quasi-geostrophic transport depends on the presence of horizontal circulations rotating about the vertical axis (eddies) along the boundary s. Such circulations can effect heat transfer without net mass transfer in any layer if T and c_{rn} are correlated so that at any height the inflow occurs at higher temperatures than the outflow. The eddy component of motion normal to the boundary is given by c_{rng}, the normal component of the geostrophic flow, and also by variations of the ageostrophic component c_{rna} about the mean $\bar{c}_{rna}$. Denoting deviations from the mean with an asterisk, the heat transfer across the boundary arising from horizontal circulations depends on the correlations $\overline{c_{rng}\, T^*}$ and $\overline{c^*_{rna}\, T^*}$. Both these terms were calculated; the second proved to be small compared with the first, as often assumed for the atmosphere, and it will be omitted.

With the foregoing definitions Eq. 1 becomes

$$\frac{H}{\rho c_w} = s \int_0^D T\, \bar{c}_{rna}\, dz + s \int_0^D \overline{T^*\, c_{rng}}\, dz \qquad . \qquad . \qquad . \qquad (2)$$

This formulation brings out the two mechanisms explicitly. The equation will next be modified to introduce non-dimensional quantities for distance and velocity as described in (I), denoted by primes. This is necessary for comparing the experiment with other general circulations such as that of the atmosphere. Given the radius of the pan r_0 and its angular rotation ω, we define $s' \equiv s/r_0$, $z' \equiv z/r_0$, $c' \equiv c/c_e$ where $c_e \equiv r_0\, \omega$, the equatorial speed of the pan. Then

$$H' = s' \int_0^{D'} T\, \bar{c}'_{rna}\, dz' + s' \int_0^{D'} \overline{T^*\, c'_{rng}}\, dz', \qquad . \qquad . \qquad (3)$$

where $H' \equiv H/c_w\, r_0^3\, \rho\omega$. In order to facilitate interpretation, the heat transfer will be presented by dividing Eq. 3 by H_0', the effective equatorial heat source, so that all fluxes will be given in per cent of H_0'. In the later sections, most equations will be given in terms of physical variables without primes to avoid an excessive number of superscripts. It should be understood, however, that all computational results are presented in non-dimensional units for ready comparison with geophysical calculations.

3. Heat balance in cylindrical coordinates

Following the route generally employed in computations for the atmosphere, we shall at first calculate the heat balance in a cylindrical coordinate system θ, r, z with velocity components u, v, w. The v-component will be taken positive toward the pole, θ positive counterclockwise about the pole. The subscript r is omitted since in cylindrical co-ordinates it will suffice to consider motion relative to the pan. In this coordinate system Eq. 2 becomes

$$\frac{H}{\rho c_w} = 2\pi r \int_0^D T\, \bar{v}\, dz + 2\pi r \int_0^D \overline{T^*\, v_g}\, dz \qquad . \qquad . \qquad . \qquad (4)$$

where $\bar{v}$ is the mean meridional wind component. From the analyses in (I) the second term on the right-hand side of Eq. 4 can be readily calculated (Fig. 1). The eddy transport is directed from heat to cold source everywhere. In low and high latitudes this mechanism

oes not carry enough heat, in middle latitudes, too much. Hence a ' direct ' mean meri-ional circulation is required near heat and cold sources, an ' indirect ' circulation in middle atitudes for balance. Now $\bar{v}$ for computing the $\bar{v}\,\bar{T}$ term, at the time of the calculations, vas known only on the top surface from direct measurement (Fig. 2). There, the $\bar{v}$ listribution is indeed qualitatively in accord with that required for heat balance. In order o determine the vertical distribution of $\bar{v}$ and the integrated transports, two approaches vere chosen :

(1) We asked what the vertical profile of $\bar{v}$ should be, assuming heat balance and mini-mum kinetic energy of the mean meridional circulation at each radius; a problem solved by Dr. G. W. Platzman (Appendix I). For this calculation no knowledge of $\bar{v}$ at any level is required, only knowledge of the heat transport required for balance (total minus eddy transport) at each radius. Hence comparison of Platzman's result with the measured distribution of $\bar{v}$ at the top (Fig. 2) furnishes an indication of the validity of the assumption. A further check is provided by comparison of the vertical motions calculated from mass continuity for the minimum meridional circulation (not reproduced) with the vertical motions obtained in (I). Both checks proved satisfactory, indicating that the meridional mass exchange actually has about the lowest kinetic energy possible for heat balance.

(2) The non-geostrophic wind components were calculated from the cylindrical equations of motion in pan coordinates :

$$\frac{du}{dt} - \frac{uv}{r} - fv = -fv_g \quad . \quad . \quad . \quad . \quad (5)$$

$$\frac{dv}{dt} + \frac{u^2}{r} + fu = fu_g \quad . \quad . \quad . \quad . \quad (6)$$

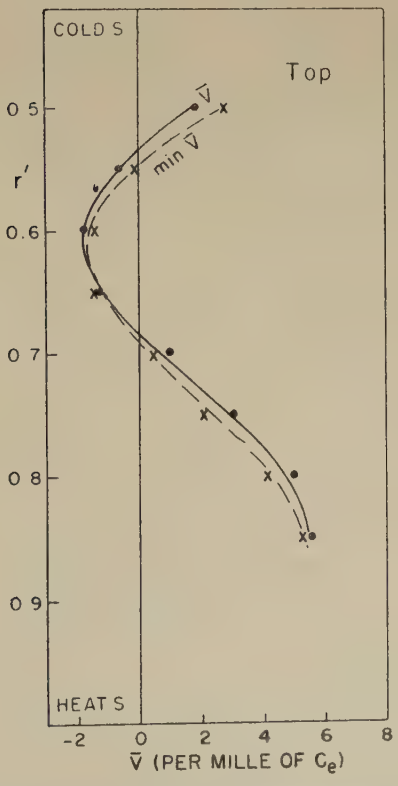

Figure 2. Mean meridional circulation at top surface as observed (solid) and as calculated from assumption of meridional circulation with minimum energy (dashed).

These equations can be evaluated in order to estimate the (non-geostrophic) $\bar{v}$'s provided we can assume that the accelerations are closely given by the geostrophic accelerations. In view of the quasi-geostrophic character of the flow field this assumption should be valid. Since steady state prevails in the coordinate system moving with the waves, $\frac{d}{dt} \equiv C_r \frac{\partial}{\partial s_r}$, where C_r is the three-dimensional velocity relative to the waves and s_r is the relative streamline or trajectory, determined by assuming isothermal motion in the interior of the liquid. C_r may be approximated by c_{rg}, the relative geostrophic velocity in the horizontal plane. This method was employed in (I) for computing the vertical motion field; justification for the procedure was given there.

Setting $u_a \equiv u - u_g$, $v_a \equiv v - v_g$, and approximating the velocity components in the centrifugal terms by the geostrophic components, Eqs. 5 and 6 become :

$$f v_a = c_{rg} \frac{\partial u_g}{\partial s_r} - \frac{u_g v_g}{r}, \quad . \quad . \quad . \quad . \quad (7)$$

$$f u_a = - c_{rg} \frac{\partial v_g}{\partial s_r} - \frac{u_g^2}{r} \quad . \quad . \quad . \quad . \quad (8)$$

Note that velocity components relative to wave coordinates appear in these equations only because $c_{rg} \frac{\partial}{\partial s_r}$ is the most convenient operator for computing individual rates of change.

Before proceeding with the calculation, we investigated whether omission of friction introduced any serious error in Eqs. 7 and 8 in the interior of the fluid. Checks were made especially in the jet-stream zone where the curvature of the horizontal velocity profiles is appreciable. It turned out, however, that the frictional retardation is one to two orders of magnitude less than the accelerations and could hence be neglected.

The meridional circulation $\bar{v} = \bar{v}_a$ was then computed from Eq. 7. This equation, however, was not applied near top and bottom of the pan where the motion is non-adiabatic (since then the estimates of individual changes fail) and where on account of unstable stratification turbulence may exist, as indeed shown by some ink trails released in the fluid. Instead, the observed $\bar{v}$ values were used at the top and the bottom distribution was obtained by extrapolation downward of vertical profiles of $\bar{v}$ drawn at intervals of $\cdot 05\ r_0$. On account of mass continuity $\int_0^D \bar{v}\, dz$ must vanish at every radius. The individual vertical profiles of $\bar{v}$ satisfied this requirement quite well; some minor adjustments for mass flow balance had to be made. Fig. 3 shows the mean meridional circulation in vertical cross-section and Fig. 4 the vertical motions computed from the equation of continuity.

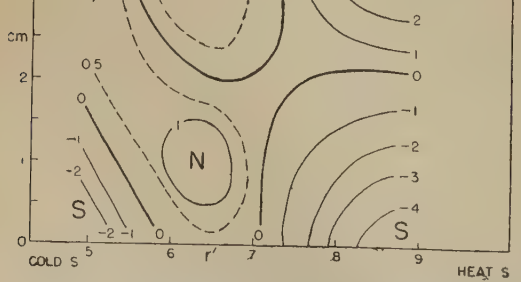

Figure 3. Vertical cross-section of the mean meridional circulation (per mille of c_e). Positive toward cold source.

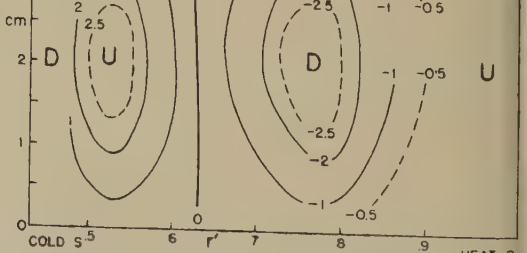

Figure 4. Vertical cross-section of vertical motion (per mille of c_e) computed from Fig. 3.

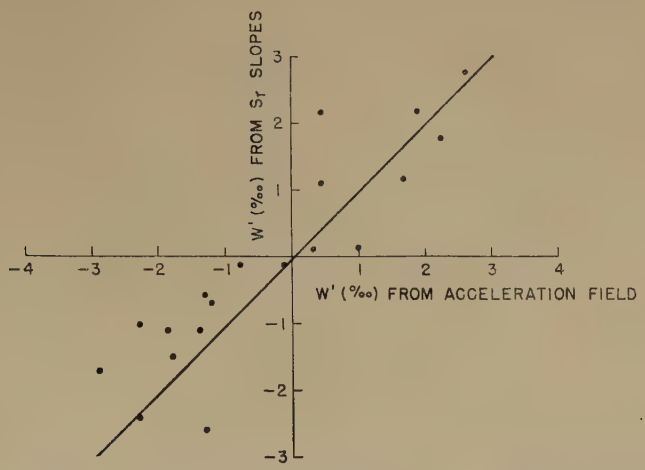

Figure 5. Correlation of vertical motion distribution in Fig. 4 and in Fig. 14(a) in (I).

$\frac{\partial}{\partial r}(\bar{v}r) + \frac{\partial w}{\partial z} = 0$. This Figure is to be compared with Fig. 14 (a) in (I). The correlation etween the two diagrams (Fig. 5) is quite satisfactory.

After all calculations had been completed, a new set of measurements was made at ·2 cm above the bottom by tellurium probe techniques (Wortmann 1953) to obtain a irect determination of the velocity field there. Comparison of $\bar{v}$ as given by the new neasurements and $\bar{v}$ for the layer 0-0·5 cm as used in the heat-balance calculation shows ood agreement of the shapes of both profiles (Fig. 6). The new data actually gave larger alues of the mean meridional circulation, but it is probable that the measurements were nade at the altitude of strongest $\bar{v}$ in the layer 0-0·5 cm and that therefore the calculated alues can be accepted with reasonable assurance. Certainly we have not over-estimated he meridional circulation.

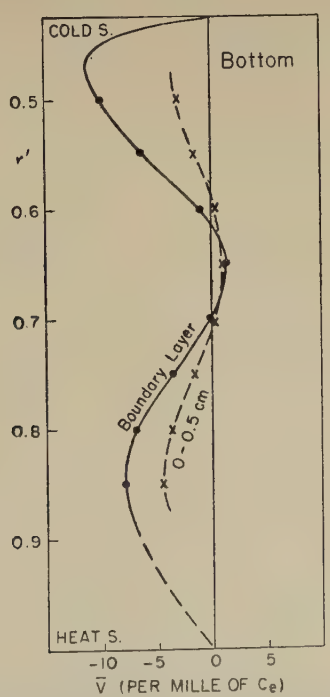

Figure 6. Mean meridional circulation as observed at 0·2 cm above bottom and calculated for the layer 0-0·5 cm.

The heat transfer computed from Fig. 3 and from Fig. 5 (a) in (I) representing th $\bar{v}\,\bar{T}$ term in Eq. 4 leads to satisfactory completion of the heat balance in Fig. 1. There th crosses indicate the total heat flow obtained from the two mechanisms. These are we within the acceptable margin of error of calculation and also within the range of error knowledge about the actual heat transport. We may conclude that the latitudinal distr bution of the relative importance of the two mechanisms of meridional heat transfer is th same as in the atmosphere, that indeed the analogy is valid to a surprising extent.

A clear picture of these mean meridional flows emerges if one introduces a Stoke stream-function $\bar{\psi}$ (Fig. 7) such that $\bar{v}r = \frac{\partial\bar{\psi}}{\partial z}$ and $\bar{w}r = -\frac{\partial\bar{\psi}}{\partial r}$ (justified by the equatio of continuity in the form $\frac{\partial}{\partial r}(\bar{v}r) + \frac{\partial}{\partial z}(\bar{w}r) = 0$). If we add independent knowledge tha strong downward motion takes place along the cold source and upward motion along th heat source, a three-cell structure of the general circulation is clearly apparent, with tw ' direct ' branches in equatorial and polar regions and a weak ' indirect ' branch in th middle, centred at the latitude of strongest westerlies. This is again in perfect agreemer with atmospheric data.

4. Heat balance in curvilinear coordinates

Since it was a principal purpose of this study to investigate the role of the jet strear in the general circulation, the preceding calculation was repeated in the curvilinea ' jet ' coordinate system (s, n, z), (see footnote p. 391). Choice of this coordinate syste was also suggested by the fact, brought out in the Introduction, that the vertical motio calculations in (I) yielded inverse results when the vertical motion was averaged arou latitude circles and along the jet axis. Fig. 8 shows the new coordinate system and th grid points used for computation. There are six lines at intervals of 0·05 r_0. The inner most of these lines, which touches the cold source, was used only for a few calculations The mean radius of each line has been entered at the right-hand edge of the diagram For purposes of comparison with the cylindrical coordinate system some results will b presented in terms of this mean radius, but in general the n-coordinate indicated alon the left-hand edge of the diagram will be used.

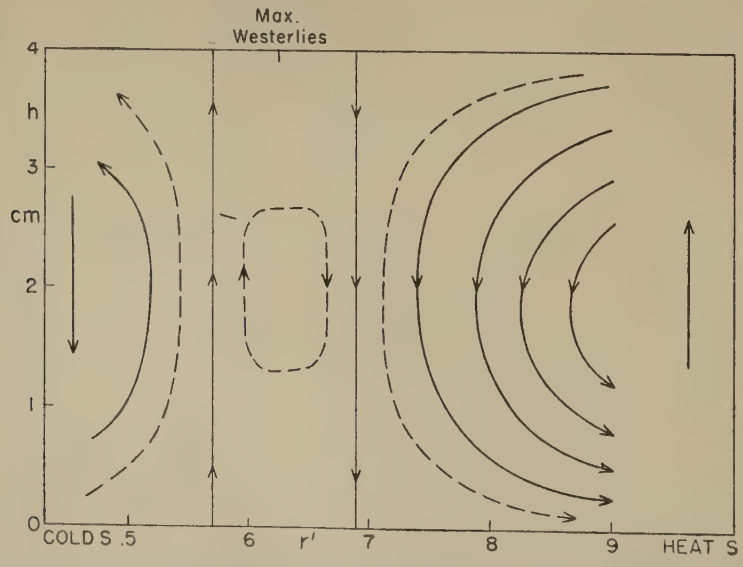

Figure 7. Stokes stream-function depicting general circulation as averaged in r-z plane. The stream functio ψ' is dimensional (cm) and is measured in units of $r_0^2\,\Omega$. The interval between full lines is $1{\cdot}0 \times 10^{-4}$ cm

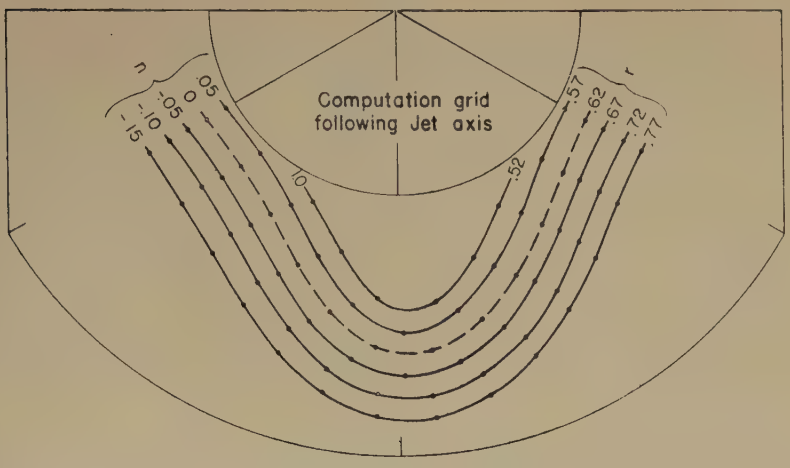

Figure 8. Coordinate system and grid of computation points fixed on jet axis.

The heat-balance computation is performed with Eq. 3, where the bar now refers to averages following the wavy lines of Fig. 8 and the asterisk denotes deviations from the average so defined. Again the eddy term is readily calculated (open circles in Fig. 9), and it is at once apparent that we encounter a situation quite different, in fact inverse, from that noted in cylindrical coordinates. The eddy transport diminishes from the heat source toward the jet axis where it becomes zero within the limits of accuracy of the calculations. *Here the ageostrophic mass circulation must execute the whole heat transport.* This outcome was by no means foreseen. We expected a somewhat larger contribution to the heat transport by the mass circulation because the coordinate system eliminates the principal eddies with respect to longitudinal averaging along latitude circles, the waves in the top layer. But most heat transport in the cylindrical coordinate system takes place in the lower levels where the vortices are located, again in good analogy with the atmosphere;

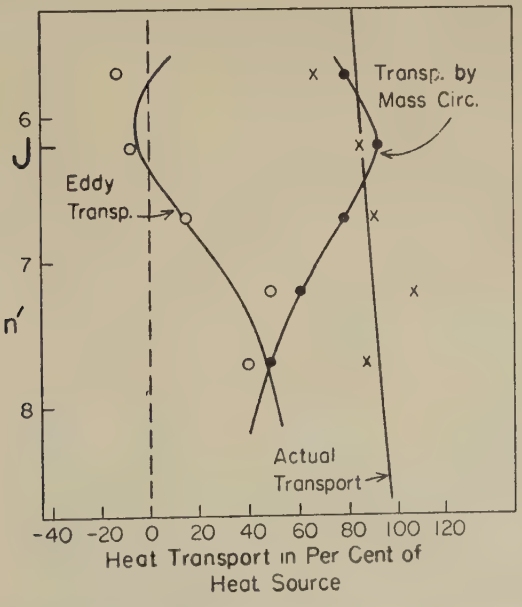

Figure 9. Heat transport in coordinate system fixed on jet axis, in per cent of equatorial heat source. Notation as in Fig. 1.

these vortices are not eliminated in the curvilinear coordinate system. If there had been a correlation $\overline{c_{rng}\, T^*}$ in the lower levels slightly higher than the correlation $\overline{v_g\, T^*}$ in cylindrical coordinates, the eddy mechanism could still have accounted for most of the heat transfer.

For direct calculation of the mass-circulation contribution the ageostrophic velocity components u_{ra} and v_{ra} in the cylindrical wave coordinate system were computed from the values of u_a and v_a obtained with Eqs. 7 and 8 using the transformations $u_{ra} = u_a + \frac{\alpha}{\omega + \alpha} u_g - \frac{\alpha^2}{2(\omega + \alpha)} r$, $v_{ra} = v_a + \frac{\alpha}{\omega + \alpha} v_g$, where α is the angular rotation of the waves, relative to the pan. These components were then projected on the lines of Fig. 8 to yield the normal ageostrophic flow c_{rna}. The above formulae allow for the fact that in the wave coordinate system, geostrophic components, and consequently also ageostrophic components, should be computed using $f + 2\alpha$ instead of f as the Coriolis parameter for the system.

After integration around each line vertical profiles of Σc_{rna} were constructed. Fig. 10 shows all computed points as well as the observed mass flow at top and bottom. In this representation the variation of the length of the s-lines is taken into account and the areas under the curves is proportional to mass flow. Along each profile the distribution of points is quite regular and nearly linear over most of the depth of the fluid, again indicating that the mass circulation has minimum kinetic energy for the required heat transport. Agreement is also good between the values interpolated for the layer 0-0·5 cm and the observed values at 0·2 cm, added later. The attempt to calculate $\bar{c}_{rna}$ near the top failed; mass balance could be obtained only by using the observed values of mass flow at the surface. This failure is not serious because the assumptions made in Eqs. 7 and 8 are not fulfilled in the top layer. The motion is non-adiabatic and the atmosphere probably exerts a marked frictional drag. Just why the discrepancy between observed and computed points increases inward has not been resolved satisfactorily.

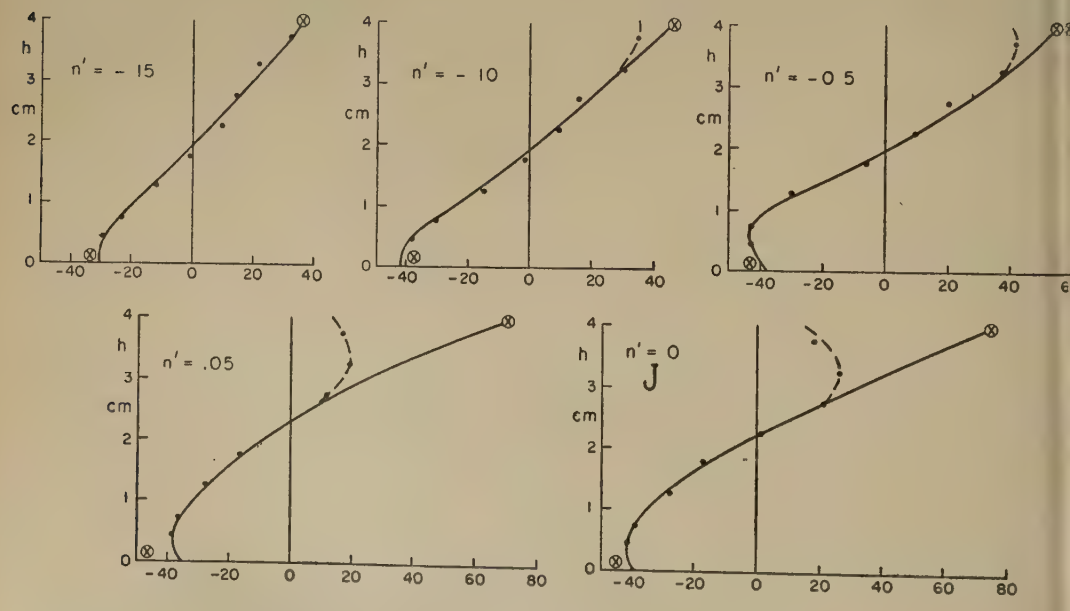

Figure 10. Vertical profiles of the computed mass circulation for the coordinate system of Fig. 8. Circles with crosses indicate measured values near top and bottom, dashed lines connect computed points near top which were rejected. Values on the scale represent the summation of the normal ageostrophic component over the grid points of Fig. 8 to take into account the variable length of the lines. Division by the number of points on each line yields the mass flow in per mille of c_e.

The mass flow varies only slowly from $n' = 0{\cdot}05$ to $n' = -0{\cdot}05$. However, the top surface values indicate a maximum mass flow at the jet axis. Beyond $n' = -0{\cdot}05$ the mass circulation decreases rapidly toward the equatorial rim. Vertical motions were calculated from the equation of continuity, which for the curvilinear coordinate system reads $\frac{1}{s}\frac{\partial}{\partial n}(\bar{c}_{rna} \cdot s) + \frac{\partial \overline{w}}{\partial z} = 0$. The result (Fig. 11) is to be compared with Fig. 15 in (I) and with Fig. 4 in this report. We note the inversion of the vertical motion distribution already found in (I). Fig. 11 and Fig. 15 in (I) agree well, strengthening confidence in the computed distribution of mass flow. It will be noted that the earlier calculation gave somewhat larger values of w. Because of the general agreement Fig. 11 has been extended inward to $n' = 0{\cdot}10$ using the values of Fig. 15 in (I) but with a reduction of the $w' = 2$ per mille line to 1·5 per mille for consistency. After this extrapolation a vertical cross-section of $\bar{c}_{rna}$ can be drawn covering large areas on both sides of the jet axis (Fig. 12).

The heat flow due to the $\bar{c}_{rna}$ mass circulation (Fig. 9) increases from the equatorial boundary of the grid toward the jet axis where it equals the total heat transport. Thus the calculation of the mass circulation is satisfactory also from the standpoint of the heat-balance requirement. The crosses in Fig. 9 show the total calculated heat transport. As in Fig. 1 verification is within the limits of acceptable error. At the highest latitude ($n' = 0{\cdot}05$) the eddy transport should be positive because of the decrease in mass-circulation transport. This result is not obtained; the computed outward eddy-heat flow, however, is very small and could easily be in error by sign.

The remarkable differences between Figs. 1 and 9 will be discussed at the end of this report. For the present it will be of interest to calculate a Stokes stream-function for the mass circulation in the curvilinear coordinate system as was done for the mean meridional circulation. For this purpose the equation of continuity can be written $\frac{\partial}{\partial n}(\bar{c}_{rna} \cdot s) + \frac{\partial}{\partial z}(\overline{w} \cdot s) = 0$. Setting $\bar{c}_{rna} \cdot s = \frac{\partial \overline{\psi}}{\partial z}$, $\overline{w} \cdot s = -\frac{\partial \overline{\psi}}{\partial n}$, the pattern of Fig. 13 results. There is now only *a single direct general circulation cell* instead of the three cells of Fig. 7. While Fig. 13 does not extend to heat and cold sources it is probable that the descent on the polar side of the jet stream merges into the descent along the cold source and that the opposite holds on the equatorial side. Thus, merely departing from the restraint of admitting only vertical circulations that are axially symmetric in a general circulation model, a representation of the general circulation by means of a single cell can be found.

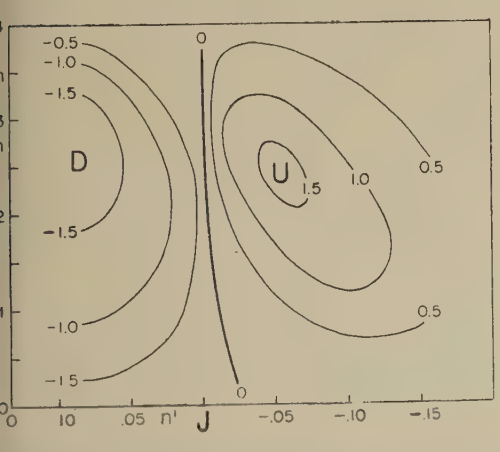

Figure 11. Vertical cross-section of vertical motion in plane normal to jet axis (per mille of c_e).

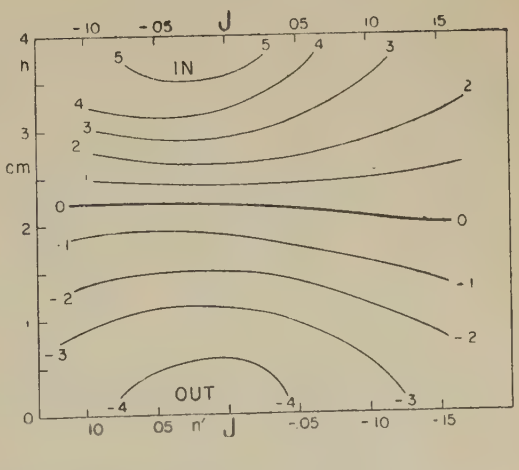

Figure 12. Vertical cross-section of mass flow in plane normal to jet axis (per mille of c_e).

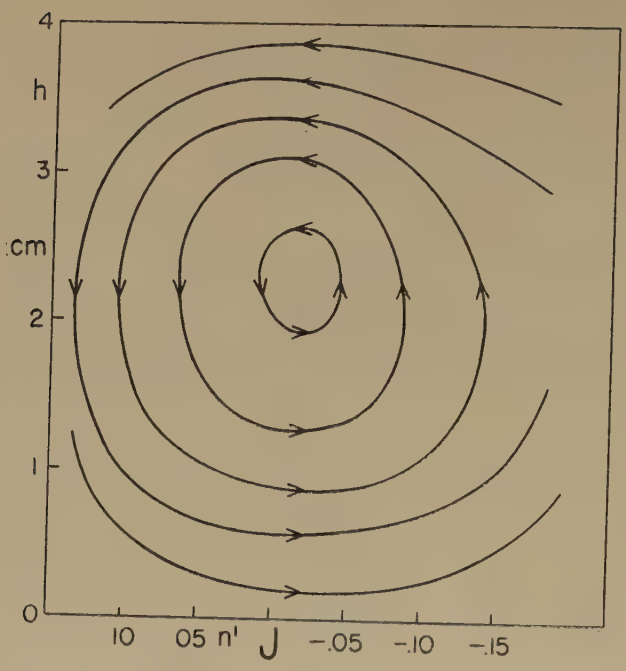

Figure 13. Stokes stream-function depicting general circulation as averaged in plane normal to jet axis. The stream-function ψ' is dimensional (cm) and is measured in units of $r_0^2\,\Omega$. The interval between full lines is $1{\cdot}0 \times 10^{-4}$ cm.

5. Angular momentum transport

For a second requirement on the general circulation the latitudinal transport of absolute angular momentum was calculated. As noted in (I) the jet stream impinges upon the cold source, hence some angular momentum undoubtedly must flow inward. Further, easterlies prevail at all latitudes at the bottom, as determined experimentally by the tellurium probes in addition to the calculation of geostrophic flow from the pressure distribution at the bottom (0 cm) shown in Fig. 7 of (I). In contrast to the atmosphere, the ground is a momentum source everywhere. The low-level vortices and the associated distribution of vertical motion do not have the function of bringing angular momentum imported from low latitudes downward to the surface by means of a correlation between u and w, which at times has been considered to be one of the main reasons for the existence of middle-latitude waves and vortices in the atmosphere. Further, the jet-stream axis is not located above a boundary (sub-tropical ridge) which separates areas where momentum is transferred to the fluid from areas where it is transferred from fluid to ground. This differs from the atmosphere where, as demonstrated by Palmén (1951), the westerly jet stream situated in the subtropics in winter lies almost exactly above the subtropical ridge. Because of all these differences we suggested in (I) that angular momentum balance requirements probably are not of primary importance in producing the observed structure of the general circulation.

Transport requirements nevertheless exist and, as in case of the heat balance, it will be of interest to see whether geostrophic eddies or an ageostrophic mass circulation accomplish the momentum transport and to compare cylindrical and jet coordinate averages. In cylindrical coordinates with absolute angular momentum $\rho\,(ur + \omega r^2)$, the transport (M) through a cylinder centred on the pole is given by

$$M = \oint_{rd\theta} \int_0^D \rho\,(ur + \omega r^2)\, vr\, d\theta\, dz \qquad . \quad . \quad . \quad (9)$$

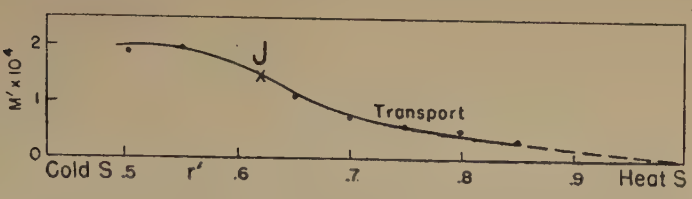

Figure 14. Transport of absolute angular momentum across latitude circles toward cold source (non-dimensional units) and transport across jet axis in coordinate system of Fig. 8 (cross).

the second term making no net contribution. Using uniform density and introducing non-dimensional distances and velocities,

$$M' = r'^2 \oint_{r'd\theta} \int_0^{D'} u'\, v'\, d\theta\, dz', \qquad . \quad . \quad . \quad (10)$$

where $M' \equiv \dfrac{M}{\rho r_0^3 c_e^2}$. The product $u'\, v'$ can next be divided as in the case of heat flow into two terms which represent momentum transport by the mean meridional circulation and by lateral eddies.

In the jet coordinate system on the other hand, the momentum transport, in physical variables, is given by

$$M = \oint_s \int_0^D \rho\,(r u_r + \omega_r r^2)\, c_{rn}\, d\,s\, dz - \oint_s \int_0^D p \sin\lambda\, r ds\, dz \quad . \quad . \quad (11)$$

Here p is pressure and λ the angle from east to the positive direction of the coordinate curve. The pressure term appears because, in the curvilinear system, the eastward acceleration due to the pressure force, integrated around the line, does not necessarily vanish; computations of the mountain effect in the atmosphere encounter the same problem.

Proceeding from these equations, the cylindrical coordinate results very closely parallel those for heat transport in Fig. 1. Fig. 14 shows the computed total transport; the general predominance of the correlation transport $\overline{u_g}\,\overline{v_g}$ agrees well with the findings of Widger (1949), Mintz (1951), and others. In curvilinear coordinates, the mass-circulation term from Eq. 11 is $\dfrac{M}{\rho} = s \int_0^D \overline{r}\,\overline{u}_r\, \overline{c}_{rna}\, dz$ which could be adequately approximated with $\overline{r}\,\overline{u}_{rg}$ replacing $\overline{r u}_r$. The cross in Fig. 14, entered at the mean latitude of the jet axis, shows the value of this term at the jet axis where it much exceeds the various geostrophic correlation contributions. It follows that the *mass circulation and the vertical shear of the zonal wind are strong enough to carry as much momentum as the correlation term in polar coordinates* in close analogy to the heat transport conclusions.

A variety of the possible correlation terms were checked in this curvilinear and the polar calculation but a detailed balance is not presented in either case because of some experimental uncertainties. There were some discrepancies with momentum-source estimates from the tellurium measurements and the pressure term in Eq. 11 turned out to be a difference of two large quantities that could not be measured with sufficient accuracy.

6. Balance of kinetic energy

In curvilinear coordinates the gain of kinetic energy per unit mass is given by dK_r/dt which for computational purposes can be approximated by $c_{rg} \dfrac{\partial K_{rg}}{\partial s_r}$ where K_{rg} denotes geostrophic kinetic energy. Averaging the gain vertically over the depth of the

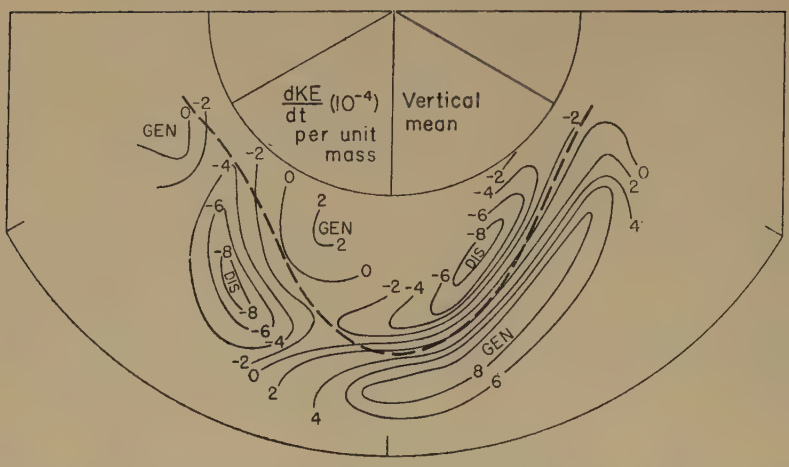

Figure 15. Distribution of actual gain and dissipation of kinetic energy, averaged vertically in non-dimensional units.

fluid the pattern of Fig. 15 results. Two regions of each sign are in evidence which correspond exactly to the areas of ascent and descent found in (I). The motion of the water through the field of vertical gradient of the horizontal velocity determines the distribution of kinetic-energy sinks and sources. One can see readily from the figure that net loss occurs in the northwest flow and gain in the southwest flow, so that trough and ridge lines separate broad regions of net loss and generation. Further, the gain east of the trough exceeds the loss to the west. All these results agree with Palmén's (1958) findings for the situation of 14 October 1954 in the United States. Alternatively, Fig. 15 also shows that net generation takes place equatorward, net loss poleward of the jet axis.

Since kinetic energy is distributed non-linearly along the vertical with a strong maximum near the top (as readily computed from Fig. 9 in I), $K_\tau \simeq K_{rs}$, the kinetic energy of the flow component parallel to the jet axis. We may therefore expect that gain and loss are also concentrated in the upper part of the fluid as corroborated by Fig. 16, where the variable length of the s-coordinate has been taken into account.

Integrating over a volume bounded by two s-lines in Fig. 8, since local time derivatives are identically zero in jet coordinates, $\int_\tau \frac{dK_{rg}}{dt}\, d\tau = \int_\sigma c_{rn}\, K_{rg}\, d\sigma$ where τ denotes volume and σ the surface bounding the volume. Introducing mean circulation and eddy terms, $\int_\sigma c_{rn}\, K_{rg}\, d\sigma = \int_\sigma \bar{c}_{rna}\, K_{rg}\, d\sigma + \int_\sigma \overline{c_{rng}\, K^*_{rg}}\, d\sigma$. These terms are readily calculated from the wind data by computing the flux through all s-lines of Fig. 8. It turns out that the eddy term is small and directed outward so that the *entire inward transport of kinetic energy is carried by the mass circulation*. This is not surprising because K_{rg} is almost constant along the s-direction in the upper part of the fluid. A consistency check from Fig. 15. proved satisfactory in that the difference in transport across two s-lines corresponds to the gain in the volume between the lines.

Kinetic energy flows everywhere from heat to cold source, with strongest transport at the jet axis (Fig. 17). Hence the kinetic energy of the jet stream is built on its equatorward margin and lost on its poleward margin. There is neither gain nor loss in a narrow volume centered on the jet axis (frictional dissipation being ignored). One cannot view the jet stream as a current that is the locus of generation of kinetic energy, which is then dispersed laterally to maintain the kinetic energy elsewhere, nor as a current that is 'driven' by outside circulations, i.e., by kinetic energy generated in polar and equatorial branches of the general circulation and carried toward the jet axis which acts as sink.

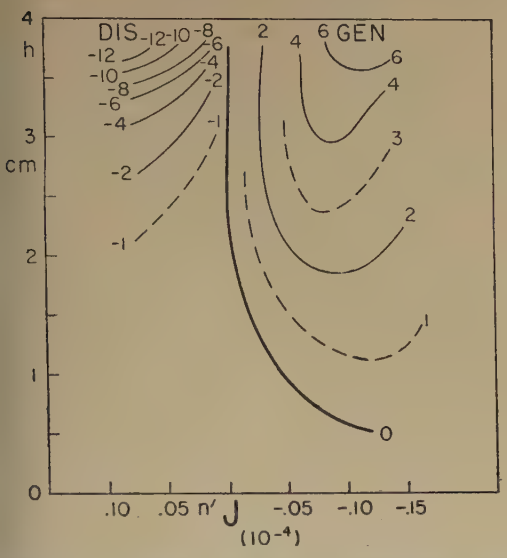

Figure 16. Vertical cross-section of generation of kinetic energy in plane normal to jet axis averaged around coordinate lines of Fig. 8. Non-dimensional units.

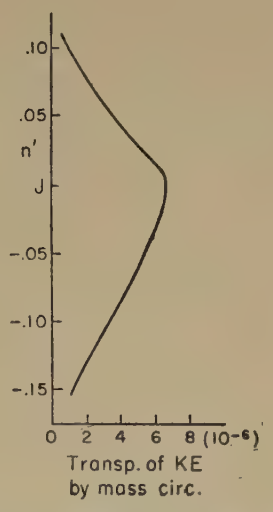

Figure 17. Transport of kinetic energy by mass circulation in plane normal to jet axis. Non-dimensional units.

The balance of kinetic energy between any two s-lines of Fig. 8 is further given from the equations of motion by

$$\int_{\sigma} c_{rn} K_{rg} \, d\sigma = -f_r \int_{\tau} (u_{ra} v_{rg} - v_{ra} u_{rg}) \, d\tau - \int_{\tau} F C_r \, d\tau, \quad . \quad . \quad (12)$$

where F is the frictional force in the direction of C_r. The first term on the right-hand side expresses the production of kinetic energy by pressure forces and the second term the total frictional work. This term can be separated into a volume integral of frictional dissipation and a boundary integral of work by friction there, since bottom and top are both moving in the wave coordinate system. The first of these effects is small over the main internal regions as stated earlier.

For calculation of the production term the flow components u, v of the polar-coordinate system were used as a matter of convenience as they were already available from the heat-balance calculation. Eq. 12 can of course be written entirely in terms of quantities in the curvilinear coordinate system. We thought it best, however, to show the form actually used for the numerical work, even though this leads to a complicated appearance of Eq. 12. In curvilinear coordinates the production may be divided into terms due to the mass circulation and due to eddies in the s-direction. The mass-circulation term is given by $f_r \int_{\tau} \bar{c}_{rna} \bar{c}_{rng} \, d\tau$ and the eddy term by $-f_r \int_{\tau} (u_{ra} v_{rg} - v_{ra} u_{rg}) \, d\tau - f_r \int_{\tau} \bar{c}_{rna} \bar{c}_{rng} \, d\tau$, the actual production less the production by the mass circulation.

Fig. 18 depicts these terms for all five volumes for which calculations could be made. Column (a) shows the actual production as expressed by the first term on the right-hand side of Eq. 12. Production is positive equatorward and negative poleward of the jet axis. The mass-circulation term is represented in column (b). Now the mass circulation flows toward lower pressure at all s-lines and latitudes. At the top, where it is directed toward the cold source, the geostrophic flow is westerly; at the bottom where it is directed toward the heat source, the geostrophic flow is easterly. In consequence there is a huge production of kinetic energy by the mass circulation. One may expect right away that this production, effective on both sides of the jet axis, is excessive and that it is one function of the distribution of direct and indirect circulation branches around the wave pattern to act as brake against this production.

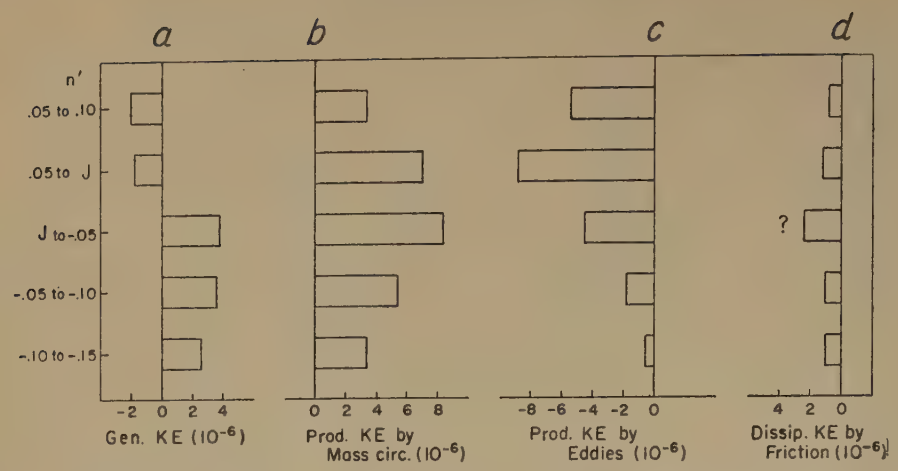

Figure 18. (a) Actual production of kinetic energy by pressure forces (non-dimensional units) in five volumes following jet axis (cf. Fig. 8); (b) production by mass circulation; (c) dissipation by eddies relative to jet axis and (d) difference between actual production and divergence of transport representing frictional dissipation.

Comparison of columns (a) and (b) of course confirms these conclusions. It follows that there must be reduction of kinetic energy everywhere by the eddies relative to jet coordinates, with maximum poleward of the jet axis (column (c)). For individual particles the braking mechanism can be seen readily for the region north of the jet axis. The amplitude of the isobars increases from the top downward. Thus particles sinking poleward of the jet axis in the southwesterly flow will move toward higher pressure at lower levels if they maintain their relative momentum over short vertical distances. The pressure force reducing kinetic energy thus acts mainly in the direction parallel to the jet axis. This reasoning does not hold for the area of sinking motion south of the jet and west of the trough. Here sinking particles must curve sharply away from the jet axis for motion toward higher pressure in the horizontal plane. It is of interest that some particles crossing the jet axis in the southwest flow will remain at the surface and therefore must lose kinetic energy while actually crossing toward lower pressure. Although a detailed analysis has not been made, it is clear that only friction can account for such changes. Friction is provided both by interaction with the atmosphere and by vertical mixing within the unstable top layer.

Column (d) of Fig. 18 finally shows the balance of Eq. 12, i.e., the excess of kinetic-energy production over divergence. It is satisfactory to note that the residual has the right sign in all five volumes. South of the jet axis the kinetic-energy production exceeds the divergence; in the north the reduction by pressure forces is smaller than the convergence. Since Eq. 12 must hold for any volume, the production term was also plotted in map form as in Fig. 15 (not reproduced). The pattern was identical with that of Fig. 15 and in all of the four centres the difference between actual generation and production by pressure forces had the right sense, i.e., there was no centre in which negative friction was computed as might easily have happened within the limitation of accuracy of all calculations.

7. Release of potential energy

As noted in (I) the energy of individual particles moving in the interior of the fluid is not conserved. Since isothermal motion may be assumed there, the internal energy remains constant. As the kinetic energy in the fluid increases with height, rising particles gain and sinking particles lose kinetic and potential energy. This produces a difficulty encountered also in the atmosphere. Transformation of potential to kinetic energy does not occur

following particles. Yet release of potential energy must be the principal means for maintaining the kinetic energy against friction. In the dishpan the problem of other energy transformations, such as of latent heat, does not arise. Moreover, large energy transformations involving potential and internal energy only, common in the atmosphere, also do not arise in the jet-stream zone of the basin. Thus the situation is somewhat simpler than in the atmosphere and, with suitable integration, one should be able to compute the release of ' free ' potential energy, i.e., that fraction of the potential-energy transformations that is used to generate kinetic energy.

Because of the steady state the total potential energy in the basin remains constant, and this holds also for any small volume fixed in the curvilinear coordinate system. Hence we can only inquire about the instantaneous rate of potential-energy release and the restoring mechanism. The condition of constant potential energy is given by $\int_\tau g\rho z \, d\tau = \text{const.}$ If $\delta/\delta t$ denotes local time differentiation in the curvilinear coordinate system, $\int_\tau gz \frac{\delta\rho}{\delta t} d\tau = 0$. From continuity $\frac{\delta\rho}{\delta t} = -\nabla \cdot \rho \vec{C}_\tau$ where ∇ is the three-dimensional gradient operator and $\vec{C}_\tau$ the three-dimensional vector wind. Then

$$\int_\tau gz \frac{\delta\rho}{\delta t} d\tau = -\int_\tau gz \, \nabla \cdot \rho \vec{C}_\tau \, d\tau = -g \int_\tau [\nabla \cdot (\rho \vec{C}_\tau z - \rho \vec{C}_\tau \cdot \nabla z] \, d\tau = 0.$$

Transforming the first term to a surface integral over the s-lines of Fig. 8

$$\int_\sigma g\rho z \, c_{rn} \, d\sigma - \int_\tau g\rho w \, d\tau = 0 \qquad . \qquad . \qquad . \qquad (13)$$

where the second term gives the potential-energy release explicitly. The physical equation for the water is

$$\rho = \rho_s (1 - \epsilon\delta) \qquad . \qquad . \qquad . \qquad . \qquad (14)$$

where ρ_s denotes standard density, $\theta \equiv T - T_s$, the temperature departure from standard, and $\epsilon = 2{\cdot}10^{-4}\ ^\circ\text{C}^{-1}$. Introducing the physical equation into Eq. 13,

$$g\rho_s \epsilon \int_\sigma Tc_{rn} \, z \, d\sigma = g\rho_s \epsilon \int_\tau Tw \, d\tau \qquad . \qquad . \qquad . \qquad (15)$$

This formulation permits calculation of the potential-energy release which, as should be expected, depends on the correlation between temperature and vertical motion. The restoring term takes the form of heat transport weighted by the altitude. For evaluation of the potential-energy release we can write at first $\overline{Tw} = \overline{T}\,\overline{w} + \overline{T^* w^*}$, where the second term denotes potential-energy transformation due to eddies along the jet stream and the averaging, as before, is with respect to the curvilinear jet coordinate system. The first term requires further inspection. When evaluation is limited to any of the five volumes between the s-lines of Fig. 8 this term may be very large owing to strong net ascent or descent. But it does not necessarily indicate tilting of the sloping isotherms with respect to the horizontal plane which is what signifies release or gain of potential energy. Ascent or descent can take place with constant orientation of the isotherms in space. To eliminate this part of the potential-energy transformations we can form means and deviations of $\overline{T}$ and $\overline{w}$ with respect to n, denoted by $\sim$ and o respectively. Then $\overline{T}\,\overline{w} = \widetilde{\overline{T}}\,\widetilde{\overline{w}} + \widetilde{\overline{T}^\circ\,\overline{w}^\circ}$ where the second term gives the tilting of the isotherms by the mass circulation explicitly (cf. also Palmén 1958). The release of free potential energy (P)

$$P = g\rho_s \epsilon \int_\tau \overline{(T^* w^* + \widetilde{\overline{T}^\circ\,\overline{w}^\circ})} \, d\tau \qquad . \qquad . \qquad . \qquad (16)$$

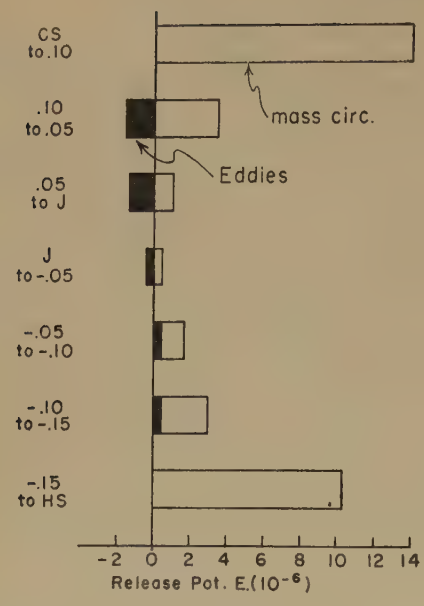

Figure 19. Release of potential energy in seven volumes (non-dimensional units). White areas: release by mass circulation; Black areas: release by eddies relative to jet axis.

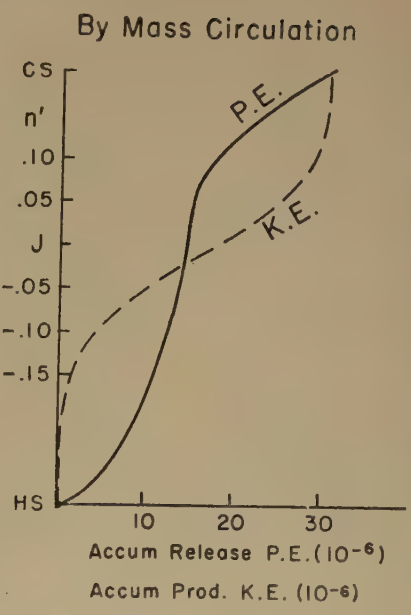

Figure 20. Release of potential energy and production of kinetic energy by pressure forces through mass circulation accumulative from heat to cold source. Non-dimensional units.

Fig. 19 shows the distribution of potential-energy release after introduction of dishpan units, i.e., after division by $\rho_s c_e^2 r_0^3 \Omega$. At first the calculation was performed only in the five volumes between the *s*-lines of Fig. 8. Since, however, the mass circulation acts to tilt the isotherms into the horizontal plane about a pivot on the jet axis, it is clear that minimum release of potential energy by the mass circulation must be encountered there, and that the release must increase progressively outward to reach maxima near heat and cold sources. Therefore the calculation was extended to both rims and, as seen in Fig. 19, it was only by means of this extension that the major fraction (about two-thirds) of the total release of potential energy was obtained. A similar extension could not be performed for the eddy potential energy; the profile of Fig. 19, however, suggests that there is likely to be a gain of one or two more units near the polar rim and no contribution near the equatorial rim. Altogether the figure shows that in the summation over the entire volume all potential-energy release is brought about by the mass circulation, whereas the eddies act to store potential energy, especially poleward of the jet axis. This storage is accomplished by ascent of relatively cold water in the northwest flow and descent of relatively warm water in the southwest flow.

The potential-energy release by the mass circulation has been plotted on an accumulative basis from heat to cold source in Fig. 20. It should be compared with the release of kinetic energy by the same circulation. Again, values are missing near heat and cold sources, but inspection of profile b of Fig. 18 indicates that only minor contributions can be expected in the border regions. After extrapolation of this profile to zero values at the boundaries, the dashed curve of Fig. 20 was obtained. Both totals are nearly identical; the potential energy released by the mass circulation is used for generation of kinetic energy by means of pressure forces in the manner of the classical simple heat-engine model sketched in Fig. 21. This description holds for the volume of the basin as a whole. When smaller volumes of the curvilinear coordinate system are considered, the energy cycle must contain a further link because most potential-energy release occurs near the rims and most kinetic-energy production near the jet axis. The intermediate link is the work done by pressure forces across the *s*-lines of Fig. 8.

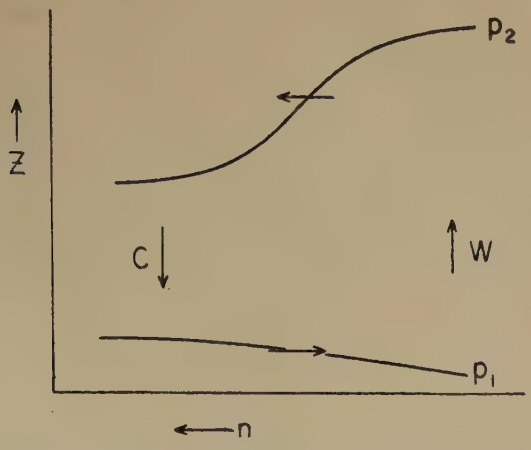

Figure 21. Model of potential-energy release and kinetic-energy production in single-cell general circulation with geostrophic jet stream. P_1 and P_2 denote two isobars.

8. The jet-stream axis

One of the previous investigations concerned with origin and maintenance of the jet stream was that of Rossby *et al* (1947), who held that since an indirect solenoidal circulation intensified a pre-existing meridional temperature gradient in a stably stratified atmosphere, one could assume the existence of such a circulation. This solution would account for the sharpness of the jet stream through development of the appropriate thermal wind. But it introduced a severe problem concerning the energy transformations, since potential and kinetic energy would increase together during jet-stream formation. Namias and Clapp (1949) and Palmén (1951) instead postulated production of the jet stream by means of a direct solenoidal circulation. This hypothesis provided potential-energy release but introduced the difficulty that because of the flattening of the temperature field by the solenoidal circulation the concentrated temperature gradient needed for quasi-geostrophic balance in the high velocity core could not be obtained or maintained around the globe. Potential and kinetic geostrophic energy would simultaneously decrease.

Superficially, the findings in this report could be cited to support either hypothesis. It should be amply clear, however, from all the foregoing that the pertinent inquiry to be made concerning the jet stream is not with respect to the maximum westerlies of the polar coordinate system but with respect to the jet-stream axis. Therefore the averaging must follow the axis. The evidence of the experiment thus favours the direct-circulation hypothesis, though in view of the thermal-balance requirement and the overproduction of kinetic energy by the mass circulation it appears unlikely that any approach, based on considerations of the mass circulation alone, can furnish an explanation of the jet stream. Nevertheless we calculated the velocity distribution in the top layer using only the mass circulation in order to see how far this approach might lead toward improved understanding of the jet stream.

For this purpose two constraints must be noted : (1). The flow is nearly geostrophic, hence a computed velocity field also determines closely the vertically-integrated temperature gradient and therewith the potential-energy release by the mass circulation. (2). The mass passing through the vertical circulation cell is subject to dynamic constraints imposed by the general laws of motion. The theorem of conservation of potential vorticity appears to be valid. As stated in (I), even the law of conservation of absolute vorticity is satisfied, with some approximation, in the top layer. From the analysis of Fig. 13, however, it is evident that some divergence and convergence do occur in the top layer and therefore we shall apply the theorem of conservation of potential vorticity in the following. At the outset at least this looks promising. The mass circulation is horizontally divergent in its

equatorial portions, convergent in its polar portions. Thus anticyclonic vorticity will b produced by this circulation in the south and cyclonic vorticity in the north ; this corres ponds to the vorticity distribution about the jet axis.

In (I) we computed the vorticity on constant-temperature surfaces in the interior o the fluid where isothermal motion can be assumed to prevail. We are now interested i the top layer (chosen from 3·5 to 4 cm) where under the experimental conditions evapora tion took place and the temperature of individual particles was not conserved. We sha therefore apply the theorem with respect to constant-pressure surfaces and, since we ar interested in the spatial vorticity distribution, we shall write the law in form of a continuit equation

$$\frac{\partial}{\partial n}\oint_s \int_0^D c_{rn}\,\zeta_{ar}\,ds\,dz = 0 \qquad . \quad . \quad . \quad . \quad . \quad (17$$

where ζ_{ar} is the absolute vorticity in the wave coordinate system. In this formulation several terms of the complete vorticity theorem are omitted. Without detailed proof, we wish to state that all these terms were computed and that although they were large in severa small areas they either cancelled or proved small in the integral over the line. As before we can write $c_{rn}\,\zeta_{ar} = \bar{c}_{rna}\,\bar{\zeta}_{ar} + \overline{c_{rng}\,\zeta^*_{ar}}$. We now set the second term aside in order to see what can be accomplished with the first term. If this omission is valid, $\zeta_{ar}\,\bar{c}_{rna}\cdot s = \text{const}$ t along the n-axis. Now $\bar{\zeta}_{ar} = f_r + \bar{\zeta}_r$, the relative vorticity. In the top layer where the flow closely follows the jet axis we may approximate the averaged relative vorticity by setting $\bar{\zeta}_r = -\dfrac{\partial \bar{c}_{rsg}}{\partial n}$. Introducing also $\bar{c}_{rna} \times s = m$, the mass circulation per unit depth,

$$\left(f_r - \frac{\partial \bar{c}_{rsg}}{\partial n}\right) m = \left(f_r - \frac{\partial \bar{c}_{rsg}}{\partial n}\right)_0 m_0 \qquad . \quad . \quad . \quad . \quad (18)$$

where the subscript will be taken to refer to observed values at the southernmost s-line in Fig. 8. Since only variations along n are considered, the partial differentials in Eq. 18 may be replaced by total differentials, and the equation can be integrated with respect to n. We obtain

$$\frac{d\bar{c}_{rsg}}{dn} = f_r - \left(f_r - \frac{d\bar{c}_{rsg}}{dn}\right)_0 \frac{m_0}{m},$$

and

$$\bar{c}_{rsg} = \bar{c}_{rsg}\Big]_0 + \int_{n_0}^{n} \left[f_r - \left(f_r - \frac{d\bar{c}_{rsg}}{dn}\right)_0 \frac{m_0}{m}\right] dn \qquad . \quad . \quad (19)$$

Fig. 22 shows the velocity distribution computed from Eq. 19 and the actual velocities. Agreement is good south of the jet axis; to the north, however, there is a large discrepancy. The mass circulation would place the jet axis close to the cold source and the core speed would be much higher than observed, requiring a larger total temperature gradient and potential energy release.* It follows that the divergence of the mass circulation suffices to explain the anticyclonic shear but that the convergence falls far short of producing the observed cyclonic shear. It also follows that a large eddy term $\overline{c_{rng}\,\zeta^*_{ar}}$ must exist to satisfy Eq. 17. Such a term was indeed computed; Fig. 23 shows the vorticity-transport balance. Formally, the eddy term arises poleward of the jet axis, because in the northwest flow with indirect solenoidal circulation, the cyclonic vorticity (shear) is much stronger than in the southwest flow with direct solenoidal circulation (Fig. 4 in I).

It appears that the principal residual question regarding the jet stream is not : ' why are jet-stream speeds so high ? ' but : ' what limits the core speed and produces the sharp cut-off on the poleward side ? ' Little more than conjecture concerning the mechanism limiting jet speed and potential-energy release can be offered at present, but there must be

* As shown by Palmén (1951) the subtropical westerly jet stream of winter in the northern hemisphere actually lies close to the northern boundary of the direct-circulation cell of the trades and not at the latitude of strongest mass circulation as in the experiment.

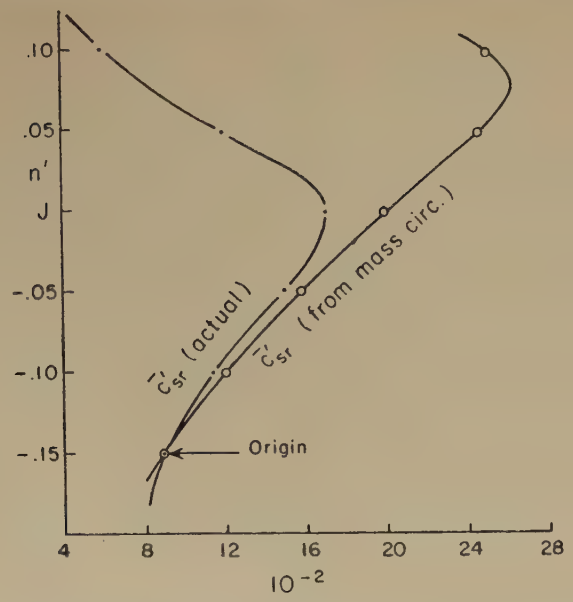

Figure 22. Profile of actual distribution of c_{sr} in layer 3·5 – 4 cm, and profile computed from Eq. 19. Non-dimensional units.

a connexion with the thermal-balance requirement. In the divergent part of the vertical circulation cell warm advection is balanced by ascent in the upper half of the fluid. Such a balance is impossible in the convergent part. As in the atmosphere a simple mass circulation would flatten the solenoid field in the absence of appropriately distributed internal heat and cold sources, and a steady jet stream with its large concentration of the lateral temperature gradient could not be achieved. Hence an eddy mechanism must maintain the temperature field ; this will be considered further in the next section.

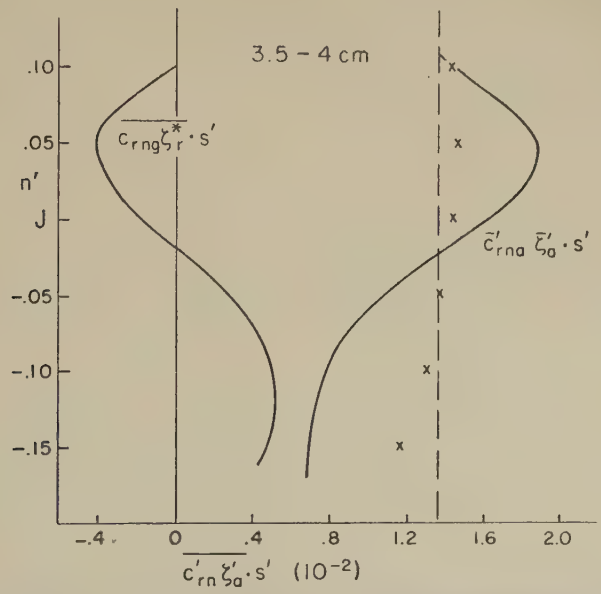

Figure 23. Vorticity transport by the mass circulation and eddies along jet axis in curvilinear coordinate system. Non-dimensional units. Crosses indicate total vorticity transport, and dashed line constant vorticity transport as required by Eq. 17.

9. The general circulation as a simple heat engine

(a) *Polar and curvilinear coordinates*

As demonstrated by the calculations of vertical motion and of heat and momentum transfer, a unique optimum way of describing the general circulation and the mechanism maintaining the circulation in the experiment has not been achieved and may not exist. Two methods of summarizing the data have been employed with different results; there may be other significant ways of representing the general circulation. Categorical statements about this circulation can be made only when the mechanisms deduced are independent of coordinate representation. When this is not the case, the limitations of the reference frame must be kept in mind in assessing the scope of permissible physical deductions. Such limitations hold for both coordinate systems studied, and a wider range of experiments must be considered to obtain more than a formal distinction.

The whole series of general circulation experiments made by Fultz (1947-1957), has shown that, depending on the rates of heating and especially of the rotation of the basin, a hierarchy of general circulations is obtained. As the rotation rate is increased at constant heating, the flow becomes more and more geostrophic owing to relative increase of the importance of the Coriolis force in the balance of forces. Three principal regimes of interest have been noted. At low rotation rates a single symmetrical trade-wind cell is observed with westerly jet stream adjoining the cold source. With increasing rotation, when the liquid is confined in an annulus (Hide 1956), this cell breaks over into steady wave patterns of the type described here and finally into unsteady circulations which, under certain laboratory conditions, take the form of a repeating cycle.

When the change from symmetrical trade-wind cell to steady wave pattern with jet stream occurs, the question arises whether this signifies a complete change in mechanism of the general circulation – replacement of circulations about horizontal axes by circulations about the vertical axis – or whether the fluid makes the minimum adjustment necessary to preserve as much of the single cell as is possible.

Hindcasting, we might have taken up this analysis with this proposition in mind: assuming that the fluid acts to preserve the single vertical cell, can we find evidence in support of this hypothesis in the observations ? *The answer evidently is affirmative.* Support for the present findings also arises in types of disturbed motion that occur very close to the breakdown region for a symmetrical trade-wind cell in an open circular disk of liquid. The disturbances characteristically take the form of one or two small-amplitude troughs spiralling toward the pole from the southwest in the upper westerlies. A corresponding (usually single) spiral of the band of strongest westerlies develops toward the pole and strongly suggests an accompanying energetically direct cell transverse to the spiralling flow (Fultz *et al* 1956). A spiral arrangement of the distribution of vertical motion with respect to the jet axis is also shown by Fig. 13 in (I). It appears that the representation of the general circulation by the coordinate system fixed on the jet-stream axis is more suggestive of the actual physical situation than the polar-coordinate representation though there is, of course, no mathematical conflict since the differences arise solely from the alteration of point of view.

The breakdown of the single cell can be explained in principle with the reasoning advanced by V. Bjerknes *et al* (1933) against axially-symmetric circulations in the atmosphere. Symmetric rings of air contracting poleward will do so under conservation of absolute angular momentum (except for frictional effects); this would require quite unrealistic wind speeds and equator-to-pole pressure gradients. This difficulty is overcome readily if one allows for a wavy or cellular pattern around the globe since this permits mass exchange over large meridional distances at relatively constant kinetic energy. Bjerknes, however, maintains the concept of a single meridional cell superimposed on the cellular pattern; his schematic diagram (Fig. 24) agrees well with the results of the present analysis.

Figure 24. Scheme of general circulation consisting of single meridional cell and cellular subtropical highs. Solid : high troposphere; dashed : surface layer. From V. Bjerknes *et al.* (1933). For comparison see Fig. 2 (b) in (I).

The meridional temperature gradient, and therewith also the potential energy release by the mass circulation, are roughly proportional to the speed of the jet stream, but the kinetic-energy production through pressure forces is proportional to the square of the wind. It is plausible, therefore, to suggest that with increasing rotation of the basin the kinetic-energy production requirement will rise more rapidly than the potential-energy release by the mass circulation, and that the breakdown occurs when the potential-energy release can no longer generate enough kinetic energy to maintain the zonal winds prescribed by the constant absolute angular-momentum profile.

(*b*) *Further comments on the heat balance*

A single symmetrical cell can remain in steady state only when the potential-energy release takes place immediately adjoining heat and cold sources, so that all vertical motion is connected with non-adiabatic processes. Heat abstraction at the cold source and heat gain at the heat source balance the instantaneous tipping of the isotherms into the horizontal plane by the mass circulation.

When wave motion replaces the symmetric cell and when the axis of strongest mass circulation follows the waves as observed, net ascent and descent will occur in the interior of the fluid where there are no effective heat or cold sources. With this, a steady temperature field can no longer be obtained with the mass circulation alone. This is especially true in the upper half of the fluid north of the centre of the cell where horizontal warm advection and descent occur together, and in the lower half south of the centre of the cell where cold advection coincides with ascent. If steady state is to be retained asymmetries having the property of being able to maintain the temperature field must arise along the axis of mass circulation. We may consider asymmetries with this property as the true ' eddies ' of the general circulation. In our case they are easily recognized as the indirect circulation in the northwest flow and the direct circulation in the southwest flow.

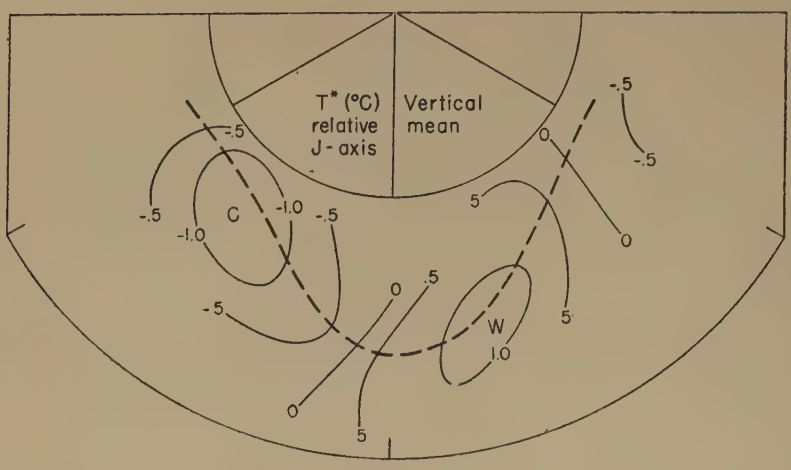

Figure 25. Temperature departures from mean in curvilinear coordinate system. Vertical averag

Fig. 25 shows the field of temperature anomaly in the curvilinear coordinate system. While the anomalies in this figure have been averaged over the whole depth of the water the pattern is nearly independent of height and may be applied at any level. Considering these temperature anomalies and the distribution of vertical motion (Fig. 13 in I) the eddy pattern represented by $\overline{T^* w^*}$ acts to remove heat downward from the upper part of the fluid north of the jet axis because the motion is upward at negative and downward at positive temperature anomalies. Further, the correlation $\overline{T^* c_{rng}}$ removes heat toward the cold source in the top layer. In this way the eddies maintain the steady temperature field, but in doing so they act to increase the potential energy (Fig. 19) and reduce the kinetic energy (column (c) Fig. 18).

These links establish a qualitative connexion between the maintenance of the thermal field and the jet stream cut-off. The mechanism for limiting the jet speed resides in the structure of the whole general circulation and particularly in the requirement for an eddy pattern north of the centre of the vertical circulation cell to counteract the release of potential energy by the mass circulation. If this interpretation is valid, the jet-stream limit is due to energetic and not dynamic factors; from the latter viewpoint no objection against actual establishment of the computed velocity profile of Fig. 22 could be raised.

Finally, the temperature anomaly pattern of Fig. 25 may be deduced assuming only a potential energy releasing mass circulation with a wavy quasi-geostrophic current at the top to avoid the problem arising from conservation of absolute angular momentum in contracting axially-symmetric rings of fluid. As stated in (I), the law of conservation of absolute vorticity is fairly well satisfied in the layer with wave motion but this, of course, is true only in the first approximation. The jet-stream axis and the axis of strongest mass circulation could not be associated in fluid, strictly obeying conservation of absolute vorticity. Fig. 12 in (I) shows that descent, increasing downward from the top to the 2 cm level takes place across the trough and that the opposite holds across the ridge. Hence there is lateral convergence around the trough in the upper part of the fluid and divergence in the lower portion, with the inverse across the ridge. From the law of conservation of potential vorticity, anticyclonic vorticity is generated in the lower half of the fluid across the trough and cyclonic vorticity across the ridge as demonstrated by Fig. 16 in (I). Since the lower current moves westward relative to the waves the highest anticyclonic vorticity occurs under the northwest inflection point of the wave pattern where the vertical motion in Fig. 12 of (I) is zero; the highest cyclonic vorticity appears correspondingly under the southwest inflection point. Fig. 7 in (I) confirms these deductions.

Now since all motion is nearly geostrophic and hydrostatic balance prevails, the foregoing analysis requires a cold column relative to the wave pattern under the northwest inflection point and a relatively warm column under the southwest inflection point, and this is what Fig. 25 shows.

(c) *The energy cycle*

Up to the 1940's the idea prevailed that the kinetic energy of the atmosphere is released in the general circulation and from there dispersed to successively smaller eddies. Along with the concept of the atmosphere as a horizontally turbulent fluid in which the transfer requirements are met by quasi-geostrophic eddies, the viewpoint has been put forward that the mean circulation is maintained by eddies which are then the locus of kinetic-energy generation. Among many writings, the work of Phillips (1956) may be cited as giving perhaps the most complete exposition of this view, backed by a machine-calculation experiment. Phillips's article also contains an extensive listing of the pertinent literature, and for this reason repetition of the entire bibliography here has been omitted.

The following notation will be used in this section, partly analogous to that of Phillips's paper and independent of the previous notation in this report.

Q	heat input integrated over volume
P	potential energy integrated over volume
K	kinetic energy integrated over volume
W	work done by pressure forces through boundaries
D	dissipation of kinetic energy by internal and skin friction
$A(P)$	advection of potential energy through boundaries
$A(K)$	advection of kinetic energy through boundaries
$\{\overline{P} \cdot P^*\}$ and $\{P^* \cdot \overline{P}\}$	conversion of mean to eddy potential energy and inverse
$\{\overline{K} \cdot K^*\}$ and $\{K^* \cdot \overline{K}\}$	conversion of mean to eddy kinetic energy and inverse
$\{\overline{P} \cdot \overline{K}\}$ and $\{P^* \cdot K^*\}$	conversion of mean and eddy potential energy to mean and eddy kinetic energy respectively.

Terms such as Q^*, $\{\overline{P} \cdot K^*\}$ and $\{\overline{K} \cdot P^*\}$ do not arise in the problem as formulated by Phillips and in this report.

Phillips used a Cartesian coordinate system which, as far as the essential definitions of $\overline{P}$, P^* $\overline{K}$ and K^* are concerned, corresponds to the polar coordinate system of this report. He summarized the energy cycle in diagrammatic form to bring out the transformations between mean and eddy potential and kinetic energy explicitly. For comparison with the dishpan we have converted his results as given in his Table 7 for days 5 to 26 to non-dimensional units (magnitude 10^{-8}) by division by $\rho_0 c_e^2 r_0^3 \Omega$, where ρ_0 is the density of the atmosphere at sea level and r_0 is taken as the radius of the earth. Fig. 26 shows the result. Because Phillips's calculations refer to an unsteady problem, the circuits on the diagram do not balance precisely. The outbound arrows from $\overline{K}$ *and* K^* denote dissipation by internal and skin friction, a distinction not made for the dishpan where we indicate the total dissipation of kinetic energy by $\overline{D}$ and D^*. The outbound arrows from $\overline{P}$ and P^* denote loss of potential energy through lateral eddy diffusivity which, as Phillips says, tends to smooth out the horizontal temperature gradients. Since ink trails released in repeated runs of the dishpan experiment give no indication whatever of the existence of such lateral diffusivity, we have considered this effect to be zero for our case.

It is the principal result of Fig. 26 that the internal energy cycle is such that $\overline{K}$ derived from K^*. Although we have not performed the necessary calculations, we presum that we would obtain the same result for the energy cycle in polar coordinates for our cas As Fig. 18 shows, the situation is quite different in the curvilinear system, and in th following we shall show the cycle for this coordinate system, using the whole volume fro $n' = -0{\cdot}15$ to $n' = 0{\cdot}10$ in Fig. 8. Since ours is an open region whereas Phillips's closed, the applicable equations are somewhat different; boundary terms occur and hea and cold source do not appear explicitly. Following are the formulae for transformatio of mean and eddy potential and kinetic energy, where the averaging now follows the j axis and where the subscript r has been omitted.

$$0 = \frac{\delta \overline{P}}{\delta t} = \{\overline{K} \cdot \overline{P}\} + A(\overline{P}) + \{P^* \cdot \overline{P}\} \quad . \quad . \quad . \quad . \quad (20$$

$$0 = \frac{\delta P^*}{\delta t} = \{K^* \cdot P^*\} + A(P^*) + \{\overline{P} \cdot P^*\} \quad . \quad . \quad . \quad (21$$

$$0 = \frac{\delta \overline{K}}{\delta t} = \overline{W} + \{\overline{P} \cdot \overline{K}\} + A(\overline{K}) + \{K^* \cdot \overline{K}\} + \overline{D} \quad . \quad . \quad (22$$

$$0 = \frac{\delta K^*}{\delta t} = W^* + \{P^* \cdot K^*\} + A(K^*) + \{\overline{K} \cdot K^*\} + D^* \quad . \quad (23$$

Not all of the terms in Eqs. 20-23 were computed directly but the equations wer solved by the residual method using the calculations discussed earlier in this report. I Eq. 20 $\{\overline{K} \cdot \overline{P}\}$ is given by the white areas of Fig. 19 from $n' = -0{\cdot}15$ to $n' = 0{\cdot}10$ the value of the term is -10 in non-dimensional units (magnitude 10^{-6}). Using Eq 13 a separate calculation was made for $A(\overline{P})$ which has a value of 7 units; therefor $\{P^* \cdot \overline{P}\} = 3$ units.

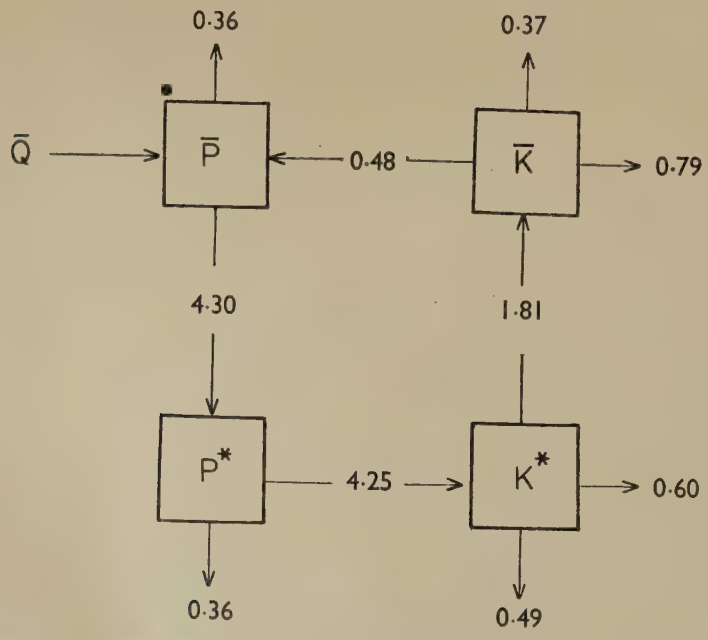

Figure 26. Energy cycle for atmosphere in Cartesian coordinates as computed by Phillips (1956) in numerical experiment. Non-dimensional units (10^{-8}). For details see text.

In Eq. 21 $\{K^* \cdot P^*\} = 3$, as given by the black areas of Fig. 19; $\{\bar{P} \cdot P^*\} = -3$ from Eq. 20 and $A(P^*)$ is therefore zero.

In Eq. 22 $\{\bar{P} \cdot \bar{K}\} = 10$ from above and $A(K) = 0$ from Fig. 17. The value of $\bar{W}$ must be 20 from Fig. 20, the difference between $\{\bar{K} \cdot \bar{P}\}$ and the total release of potential energy or production of kinetic energy by pressure forces through the mass circulation. Since we are not interested in distinguishing between $\bar{D}$ and D^* for the dishpan, we shall assume $\bar{D} = 0$, so that $\{K^* \cdot \bar{K}\} = -30$. In the other extreme case ($D^* = 0$) we would have $\{K^* \cdot \bar{K}\} = -23$, retaining the same sign for this term which is the feature of importance.

In Eq. 23 we already have $\{P^* \cdot K^*\} = -3$ and $A(K^*) = 0$ from the discussion of the kinetic-energy balance. The term $\bar{W}^*$ must be the difference between column (c) of Fig. 18 (-23 units) and $\{P^* \cdot K^*\}$, hence $W^* = -20$ units. Using $\{\bar{K} \cdot K^*\} = 30$, $D^* = D = -7$, which checks with column (d) of Fig. 18. It is of interest that the total work across the boundaries $W = \bar{W} + W^* = 20 - 20 = 0$ so that energy balance is obtained by $A(P) + D = 7 - 7 = 0$.

Following is a summary of the numerical results, using the order of Eqs. 20-23.

$$0 = \frac{\delta \bar{P}}{\delta t} = -10 + 7 + 3 \quad . \quad . \quad . \quad . \quad . \quad (20a)$$

$$0 = \frac{\delta P^*}{\delta t} = 3 + 0 - 3 \quad . \quad . \quad . \quad . \quad . \quad . \quad (21a)$$

$$0 = \frac{\delta \bar{K}}{\delta t} = 20 + 10 + 0 - 30 - 0 \quad . \quad . \quad . \quad . \quad (22a)$$

$$0 = \frac{\delta K^*}{\delta t} = -20 - 3 + 0 + 30 - 7 \quad . \quad . \quad . \quad . \quad (23a)$$

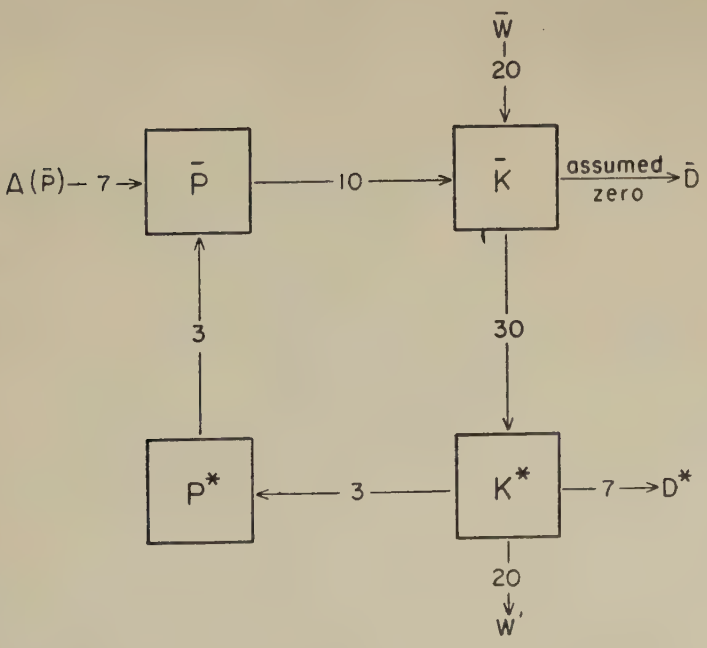

Figure 27. Energy cycle for experiment in curvilinear coordinates. Non-dimensional units (10^{-6}). For details see text.

The numbers are a couple of orders of magnitude larger than those of Fig. 26, showing that scale equivalence in space and time between dishpan and atmosphere does not necessarily lead to equivalent magnitudes of the energy transactions.* Fig. 27, however, which presents the results of Eqs. 20-23 in diagram form, shows the essential point : the energy cycle has the inverse sense of that of Fig. 26. All kinetic-energy production resides with the mass circulation, and the eddies assume a braking role throughout in support of the classical viewpoint that generation of kinetic energy takes place in the largest circulations and is then transformed to energy of the eddies superimposed on the mean circulation.

10. Conclusions

Like many studies, ours has raised more questions than it has answered. Outstanding among other problems is the explanation of the jet-stream cut-off, the explanation of shape and intensity of the mass-circulation cell of Fig. 13, and a theoretical deduction of the observed vertical and lateral temperature gradients. Without further specification of quantities such as the heat transfer (H) as input data it is difficult to extend the analysis. Such specification, however, would hardly be significant. The fluid has some freedom in selecting the amount of heat to be transmitted, so H cannot be assumed to be prescribed by experimental conditions.

A new experiment has been run in which the rotation rate is higher than that of the present case. This experiment yielded long waves and a jet stream with a cyclical variation, so that the steady-state concept must be applied to the cycle rather than to the instantaneous circulations. It is hoped that introduction of the time variable will provide further information about the constraints on the general circulation, and on the jet stream.

Acknowledgments

In the first place, the authors wish to express their thanks to Dr. G. W. Platzman who participated extensively in discussions of the material throughout the entire study and who helped especially in the incorporation of the discussion of the energy cycle in the last section. They also wish to acknowledge suggestions made by Dr. E. Palmén, Dr. J. S. Malkus and Dr. C. H. B. Priestley, and the assistance of their staff, especially Mr. R. Kaylor, Mr. T. Krishnamurti and Mrs. J. Weil. The research was carried out with the support of the Office of Naval Research, United States Navy, and the Geophysical Research Directorate, U.S. Air Force.

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* Presumably this is related to the large differences in thermodynamic efficiency in the two systems, and to differences in the thermodynamic pressure work term representing conversion from thermal to mechanical energy in the thermodynamic energy equation. Fultz *et al* (1956) have shown that this term, $p\nabla \cdot \vec{C}$, when written in non-dimensional form consistent with the variables of this paper, must be multiplied by $g\epsilon D/2c_v$ where c_v is specific heat at constant volume. This factor has a value of about 0·2 in the atmosphere and 10^{-8} in the experiment. This is probably the source of the difference in magnitudes between Figs. 26 and 27.

Appendix I

Profile of a minimum meridional flow $\overline{V}'$

Suppose we regard the temperature field as fixed and the northward heat transport by the mean meridional flow H_m' as given at a particular radius by subtracting the observed eddy flow (eddy deviations from longitude average at fixed r) from the required total transport. We ask what profile of $\overline{V}'$ against z will accomplish the transport H_m' with minimum meridional kinetic energy i.e., minimum $\overline{V}'$ in the mean squared sense with respect to a vertical average.

Let $\widetilde{(\ )} \equiv \frac{1}{D}\int_0^D (\ )\, dz$ where D is the depth of the fluid. The conditions arising from the above together with mass continuity are

(1) $\widetilde{\overline{V}}' = 0$

(2) $\widetilde{T\overline{V}}' = H_m' K$ where K is a constant yielding the correct dimensions

(3) $\widetilde{\overline{V}'^2} =$ minimum

Variation of (1), (2), and (3) yields

(4) $\widetilde{\delta\overline{V}}' = 0$

(5) $\widetilde{T\,\delta\overline{V}'} = 0$ (since we do not consider variations of the T field)

(6) $\widetilde{\overline{V}'\,\delta\overline{V}'} = 0$

(6) being the necessary condition for minimum energy. For any choice of $\delta\overline{V}'$ consistent with (4) and (5) we see that (6) will be satisfied if

$$\overline{V}' = aT + b,$$

where a and b are constants independent of z. For, with this choice

$$\widetilde{\overline{V}'\,\delta\overline{V}'} = \widetilde{(aT+b)\,\delta\overline{V}'} = a\widetilde{T\,\delta\overline{V}'} + b\widetilde{\delta\overline{V}'} = 0 + 0.$$

From (1), $\widetilde{\overline{V}}' = 0 = \widetilde{aT+b} = \widetilde{aT} + b$

so $b = -a\widetilde{T}$

and $\overline{V}' = aT + b = a(T - \widetilde{T}) = aT^\circ$

where T° denotes the deviation from the vertical average. From (2),

$$KH_m' = \widetilde{T\overline{V}}' = \widetilde{T^\circ\overline{V}}' \quad \text{since} \quad \widetilde{\overline{V}}' = 0$$

$$KH_m' = \widetilde{T^\circ aT^\circ} = a\widetilde{T^{\circ 2}}$$

$$a = \frac{KH_m'}{\widetilde{T_0^2}}$$

Thus we have $$\overline{V}' = KH_m' \frac{T^\circ}{\widetilde{T_0^2}}$$

In the experiment, since T and T° are nearly linear functions of z this implies that the minimum profile is one of $\overline{V}'$ linear in z.

The foregoing procedure was suggested by Dr. G. W. Platzman.

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The convective regime and temperature distribution above a horizontal heated surface

By J. F. CROFT
Royal Military College of Science, Shrivenham, Wiltshire

(Communicated by Sir Graham Sutton, F.R.S. Manuscript received 30 January 1958)

SUMMARY

The convective flow and temperature distribution above a horizontal heated surface in air have been examined by optical methods. Attention has been given both to the open system and to enclosed air layers. The observations show the existence close to the surface of a mode of convection similar to the cellular convective motion previously noted in shallow enclosed layers. This motion is found to be present both in the open system and in the lower part of an enclosed layer of wide separation, i.e., large Rayleigh number.

The mean temperature profile in this region has been found for a range of surface temperatures and the results show that the logarithm of the mean absolute temperature (T) varies as the power ($-\frac{1}{2}$) of the height (Z) above the heated surface. This measured distribution has been compared with theoretical predictions in particular that of the Priestley similarity theory, which properly refers to convection above an infinite surface and in which the postulated conditions will apply more exactly in higher layers. Nevertheless the measured temperature profile in the surface region is of the form predicted by the Priestley theory with the experimental value ($-\frac{1}{2}$) in place of the theoretical ($-\frac{1}{3}$) as the power of the height (Z).

1. INTRODUCTION

The conditions in the unstable fluid above a horizontal heated surface, with or without a parallel cooled upper boundary, have been studied theoretically and experimentally by many investigators, without any clear agreement on the temperature profile in the fluid, or on the conditions which govern the various observed changes in the convective regime.

In particular recent experiments reported by Thomas and Townsend (1957) do not fully support the predictions for the temperature profile of theories by Malkus (1954) and by Priestley (1954). As a part of a programme of work on convection in these conditions measurements of the temperature profile in the air immediately above a heated plate have been made by an interferometric method, and the flow patterns have been studied visually and photographically by a direct shadow method. These observations give a clear indication of the flow condition and temperature profile in the air layers close to the surface.

2. THEORETICAL PREDICTIONS AND THEIR VERIFICATION

(a) *Conditions above an open surface*

Priestley's treatment of the problem assumes that the heating of a large horizontal surface in an ocean of fluid will throw up a succession of buoyant elements which mix by small-scale turbulence with the surrounding fluid. By similarity arguments, supported by a dynamical treatment, it is deduced that the logarithm of the mean absolute temperature T of the fluid should vary as a power of the height Z above the surface

i.e. $$-\frac{g}{T}\frac{\partial T}{\partial Z} = -g\frac{\partial}{\partial Z}(\ln T) = A Z^{-\delta} \quad . \quad . \quad . \quad (1)$$

where $\delta = 4/3$; A is a constant; g is the acceleration due to gravity.

Hence $\ln T = \dfrac{A}{(\delta - 1)g} Z^{1-\delta} + C$, where C is a constant

i.e., $$\ln T = B Z^{(1-\delta)} + C \quad . \quad . \quad . \quad . \quad (2)$$

where $B = \dfrac{A}{(\delta - 1)g}$ i.e., a constant.

The Priestley theory gives a value of $\delta = 4/3$, but Thomas and Townsend, following an argument by Malkus, suggest that $\delta = 2$ and hence that $\ln T$ is proportional to Z^{-1}. They claim that their own measurements of the temperature profile in a system with enclosing walls, using a fine-wire resistance thermometer, support this power law but do not satisfy the complete predictions of the Malkus theory. It seems desirable therefore to see whether independent experimental data can be fitted to a relation of the form of Eq. (2).

If the temperature at an infinite height above the heated surface be denoted as T_A, then in Eq. (2) as $\delta > 1$, $C = \ln T_A$, and

$$\ln \frac{T}{T_A} = B\, Z^{(1-\delta)} \quad . \quad . \quad . \quad . \quad . \quad (3)$$

If, as in the Priestley similarity theory, the convection process is assumed to be of such a form that its consequences are independent of the molecular constants of the fluid, i.e., thermal conductivity and viscosity, then within this region it may be shown that

$$\ln \frac{T}{T_A} = D\, Q^{2/3} g^{-\frac{1}{3}} Z^{-\frac{1}{3}} \quad . \quad . \quad . \quad . \quad (4)$$

where $Q = \dfrac{H}{\rho\, C_p\, T}$ is a function of the heat flux H

D is a constant which may depend on the Prandtl number $\dfrac{\nu}{K}$

ν is the kinematic viscosity of the fluid

K is the thermal diffusivity of the fluid

ρ is the density of the fluid

C_p is the specific heat at constant pressure of the fluid.

As pointed out by Thomas and Townsend this relation may be expressed in a non-dimensional form by the introduction of two convenient parameters θ_0 and Z_0 such that

$$\theta_0 = \left(\frac{Q^3}{Kg}\right)^{\frac{1}{3}} \quad \text{and} \quad Z_0 = \left(\frac{K^3}{Qg}\right)^{\frac{1}{3}}$$

Eq. (4) then becomes

$$\theta_0^{-1} \ln \frac{T}{T_A} = D \left(\frac{Z}{Z_0}\right)^{-\frac{1}{3}} \quad . \quad . \quad . \quad . \quad (5)$$

These authors have made the measured temperature profiles for three different surface temperatures conform to a non-dimensional relation of this kind by choosing appropriate values of T_A and suggest that their results are best described for values of Z greater than $2Z_0$ by the relation

$$\theta_0^{-1} \ln \frac{T}{T_A} = 3{\cdot}0 \left(\frac{Z}{Z_0}\right)^{-1} \quad . \quad . \quad . \quad (6)$$

It is this power law which they suggest is consistent with the Malkus theory, which requires however, a constant 2·0 in place of the experimental value 3·0 of Eq. (6). Their temperature measurements were made using a fine-wire resistance thermometer which required correction for heat conduction along its supports and presumably for radiative exchange with the surroundings. The measurements were made above a heated surface which comprised the base of a roughly cubical box.

(*b*) *Conditions in enclosed layers*

It is also necessary to consider the conditions which apply in enclosed air layers. Experiments by Bénard (1900) show that in the case of shallow horizontal layers, when the temperature gradient exceeds a certain critical value convective motion begins in a regular

pattern. The layer divides up into a pattern of identical cells in which the fluid is circulated in a rotary motion, known as cellular convective motion. The work of Rayleigh (1916), Jeffreys (1926 and 1928) and Pellew and Southwell (1940) establishes that the nature of this process at any stage depends on the value of the non-dimensional Rayleigh number (Ra), defined by : $Ra = -\frac{\beta g d^4 \alpha}{K\nu}$

where β is the temperature gradient in the pre-convective state, assumed constant

α is the coefficient of expansion of the fluid

d is the vertical height of the fluid layer.

Sustained convection is possible only if this quantity exceeds a definite critical value Ra_{crit} and the value appropriate to various boundary conditions has been calculated to be : two free plane surfaces, 658; one free, one rigid plane surface, 1101; two rigid plane surfaces, 1708.

Furthermore, in the case of two rigid plane surfaces it has been found experimentally that for gases the cellular motion is in the form of hexagonal or tetragonal prisms with cooled gas descending at the centre and warm gas ascending at the boundaries. In air it is found that the cellular motion transforms into turbulence when Ra exceeds the value 45,000 – see De Graaf and Van der Held (1953).

3. Experimental procedure for the temperature measurement

(*a*) *General*

The heated surface was that of a smooth aluminium plate, accurately levelled and electrically heated from below. The surface temperature of the plate was measured by four copper-constantan thermocouples embedded below the surface. These also showed that an even distribution of temperature was maintained over the surface. An extra thermocouple was placed below a layer of hardboard of known thermal conductivity located under the aluminium plate. This allowed an estimate to be made of the heat transference from the surface. The size of the surface was limited to a 10 × 8 cm section in order to accommodate the apparatus in one arm of a small Michelson interferometer which was adapted pending the construction of a large Mach-Zehnder interferometer. The whole assembly was protected from stray draughts by a large transparent enclosure with suitable access for the light beam from the optical system used for visual observation of the air flow. The ambient and room temperatures were measured by sub-standard mercury-in-glass thermometers, one of which was located where the rising air emerged from the transparent enclosure some 50 cm above the plate surface.

(*b*) *The interferometric measurement of temperature*

An interferometric method of temperature measurement, whilst introducing certain problems of its own, removes the need for the radiation and conduction corrections of a material thermometer. It also eliminates the possibility of the phenomena under observation being influenced by the insertion of the thermometer itself. It readily yields a spatial mean temperature over the complete horizontal section at any level in the field and allows the analysis of the mean temperature profile to be pursued right down to the heated surface.

A system of monochromatic fringes was obtained at right angles to the unheated surface by aligning the fringes with one edge of an accurately machined metal block included in the field of view. The temperature distribution above the heated surface was found by measuring the resultant displacement of these fringes.

If n denotes the refractive index of air having density ρ, then $(n-1)$ is very nearly proportional to ρ. Comparing conditions at an arbitrary height Z above the plate with those (n_p, ρ_p) adjacent to the plate

$$\frac{n-1}{\rho} = \frac{n_p - 1}{\rho_p} = \frac{n_p - n}{\rho_p - \rho}$$

so that

$$n_p - n = (n_p - 1)\left(1 - \frac{\rho}{\rho_p}\right) \quad . \quad . \quad . \quad (7)$$

If the total effective path length through the disturbed air is L, and this length contains in vacuo, N_0 wavelengths λ_0, whilst there are N wavelengths λ at absolute temperature T, and N_p wavelengths λ_p at the absolute temperature T_p of the plate, then

$$L = N_0 \lambda_0 = N \lambda = N_p \lambda_p$$

$$n_p = \frac{N_p}{N_0} \qquad n = \frac{N}{N_0}$$

The fringe shift between the plate ($Z = 0$) and the height Z is then

$$\begin{aligned} \Delta N &= N_p - N \\ &= N_0 (n_p - 1)\left(1 - \frac{\rho}{\rho_p}\right) \\ &= \frac{L}{\lambda_0}(n_p - 1)\left(1 - \frac{T_p}{T}\right) \end{aligned}$$

from which it is found that

$$T_p - T = T_p \frac{\Delta N}{\Delta N - (n_p - 1)\dfrac{L}{\lambda_0}} \quad . \quad . \quad . \quad (8)$$

The path length L was nominally twice the length of the plate (i.e. 16 cm) but was subject to a correction discussed below, λ_0 was that of the Sodium D lines and was taken as 5893×10^{-8} cm, and $(n_p - 1)$ was calculated from the value for air of 0·000292 at S.T.P.

The fringe systems were recorded photographically and examined by projection on to a screen with a magnification of some 40 times. The fringe shift ΔN at each height Z was calculated as the mean for all fringes in the field of view (usually about ten), each individual estimate being made to 1/20th of a fringe, corresponding to an uncertainty in temperature of $\pm$ 0·2 °C.

(c) *The end correction*

As the working space in the interferometer arm was limited, a guard ring could not be added to the system to give a more exact definition to the disturbance above the surface. The total observed fringe shifts in a beam passing over the centre of the plate thus include the contribution of marginal regions in which the temperature field is distorted by convection from the sides of the plate and lateral conduction in the air. The amount of this contribution was assessed by a step-by-step evaluation of successive elements of the light path crossing the marginal regions, each computed from observations of the fringe shifts produced in beams passing along the edge of the system, i.e., wholly within these marginal regions. A composite correction was obtained in this way at all heights for varying values of plate temperature. It was found that a convenient procedure was the adoption of a ' corrected length ' L, varying with the plate temperature T_p, which by inclusion in Eq. (8) gave corrected values of T.

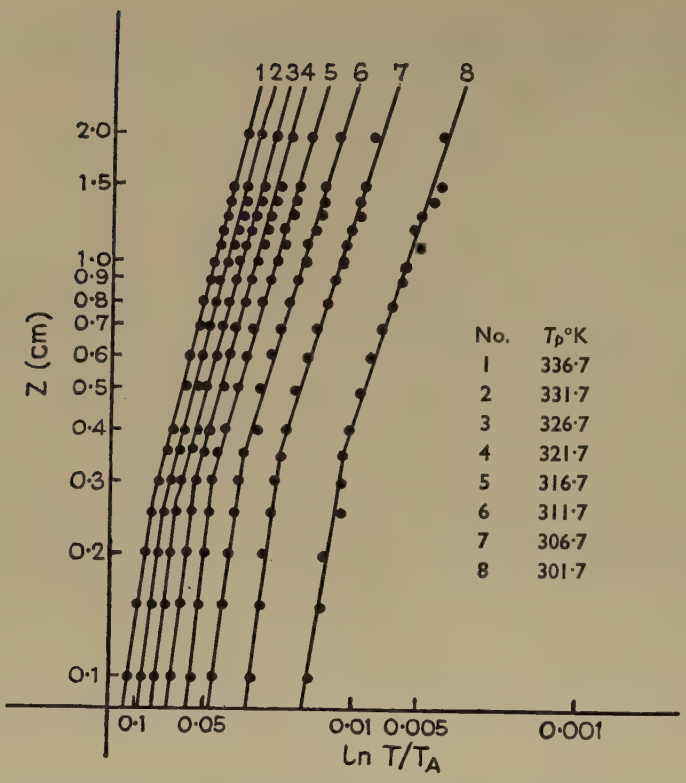

Figure 1. The logarithmic plot of Z against ln T/T_A for various surface temperatures T_p.

4. Experimental results

As there were fluctuations in the temperature in the field at any given point an extensive series of measurements was made to obtain reliable mean values of temperature. Average values of the mean air temperature T for varying heights Z were compiled for selected values of the surface temperature T_p. Fig. 1 shows a logarithmic plot of the height Z against $\ln \frac{T}{T_A}$, where T_A refers to the temperature at an infinite height above the surface. In the present case, T_A was taken as the temperature at the exit from the transparent enclosure. This gives a well-defined characteristic temperature which differs only little from the limiting value.

From Fig. 1 it can be seen that, for the temperature and height range considered, the temperature profile may in each case be analysed into two parts. For each surface temperature there is a certain height Z above which the distribution follows closely Eq. (3), i.e., $\ln \frac{T}{T_A} = B\, Z^{1-\delta}$ where B and δ take the values shown in Table 1. Though it may be seen that the index δ decreases with increasing surface temperature, the results can be satisfactorily represented within the experimental limit by the use of the mean value $\delta = 1{\cdot}5$

i.e.,
$$\ln \frac{T}{T_A} = B\, Z^{-\frac{1}{2}} \quad . \quad . \quad . \quad . \quad . \quad (9)$$

The constant B increases virtually linearly with increasing surface temperature.

TABLE 1. EXPERIMENTAL VALUES OF B AND δ

Plate temperature T_p (°K)	Height Z above which Eq. (3) holds (cm)	$B \times 10^3$	δ
336·7	0·25	48	1·4
331·7	0·25	41	1·4
326·7	0·25	35	1·5
321·7	0·3	30	1·5
316·7	0·35	24	1·6
311·7	0·35	18	1·6
306·7	0·4	12	1·6
301·7	0·4	6	1·6

In immediate contact with the surface there is a region a few millimetres thick in which the temperature profile differs from that above. The extent of this region decreases with increasing surface temperature, and though Eq. (3) using a value for δ of 1·2 can describe the temperature profile quite well, the best agreement is obtained by an equation of the form

$$\ln \frac{(T - T_A)}{(T_p - T)} = - m Z \quad . \quad . \quad . \quad . \quad (10)$$

The value of m increases with increasing plate temperature.

To see whether the measured temperature profiles conform to the general non-dimensional form of Eq. (6) it is necessary to introduce the kinematic heat flux Q, contained in the non-dimensional quantity θ_0. This has been done by the use of an expression, verified by Thomas and Townsend and consistent with the measured heat transfers, namely

$$\ln \frac{T_p}{T_A} = 3{\cdot}4 \left(\frac{Q^3}{Kg}\right)^{\frac{1}{4}} = 3{\cdot}4\, \theta_0 \quad . \quad . \quad . \quad (11)$$

As these authors point out, this expression is very similar to the heat transfer in enclosed air layers at high Rayleigh numbers, found by Malkus and others. The visual observations made of the flow regime in the two cases, described later in this paper, strongly indicate that this similarity is to be expected.

Values of θ_0 and Z_0 were calculated for the appropriate plate and ambient temperatures and these values are shown in Table 2. Fig. 2 shows a plot of $\theta_0^{-1} \ln \frac{T}{T_A}$ against Z/Z_0. It may be seen that for this temperature range, for values of Z/Z_0 greater than 2·2, the temperature distributions can be represented very well by the non-dimensional relation

$$\theta_0^{-1} \ln \frac{T}{T_A} = 3{\cdot}1 \left(\frac{Z}{Z_0}\right)^{-\frac{1}{4}} \quad . \quad . \quad . \quad . \quad (12)$$

TABLE 2. CALCULATED VALUES OF θ_0 AND Z_0

T_p (°K)	T_A (°K)	θ_0	Z_0
336·7	297·9	0·039	0·123
331·7	297·8	0·035	0·127
326·7	297·6	0·030	0·129
321·7	297·5	0·025	0·136
316·7	297·3	0·020	0·143
311·7	297·2	0·015	0·155
306·7	297·0	0·010	0·173
301·7	296·9	0·005	0·215

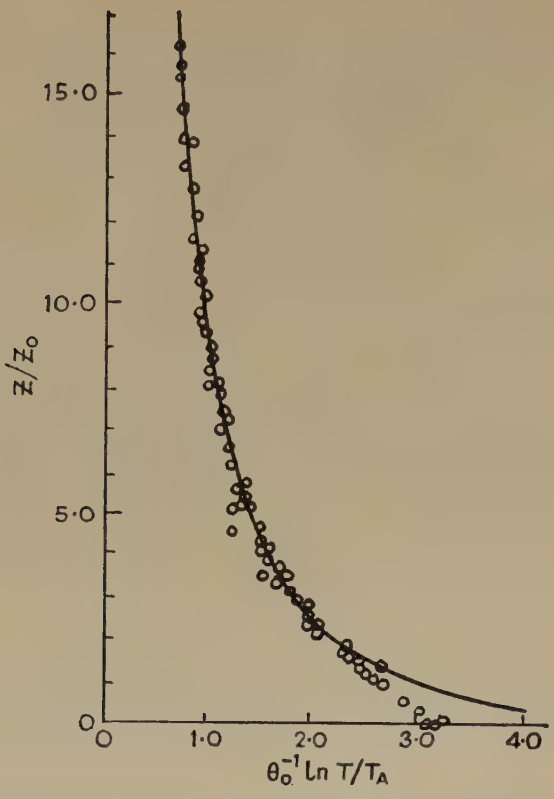

Figure 2. The experimental points shown are those of all the curves in Fig. 1. The full line represents the equation $\theta_0^{-1} \ln \frac{T}{T_A} = 3{\cdot}1 \left(\frac{Z}{Z_{\text{II}}}\right)^{-\frac{1}{4}}$.

Eqs. (9) and (12) both show the existence of a region in the fluid, which begins a few millimetres above the surface, in which the temperature decay is such that ln T is proportional to $Z^{-\frac{1}{4}}$. This decay law may be compared with the theoretical predictions of Priestley and Malkus which suggest powers of Z of $-\frac{1}{3}$ and -1, i.e., values of δ in Eq. (3) of 4/3 and 2 respectively. Neither of these are in exact agreement with the measured distributions, which give a mean value of $\delta = 3/2$, but the Priestley theory conforms the more closely, particularly at the higher temperatures where δ tends to be a little less than 3/2.

Though the Priestley theory is concerned with the heating of a large horizontal surface, conditions very close to the surface in the present investigation should be closely analogous. It is worth seeking, therefore, visual evidence of the process postulated by Priestley, namely the continuous mixing of buoyant elements with their environment by turbulence on a smaller scale.

Furthermore it is found by calculation based on the measured temperature distributions that radiation and conduction at the surface can account for only about one-third of the observed heat transfer. The remainder must be due to motions in the fluid and it would appear that a significant circulation of fluid must be present close to the surface. Again visual observation of the fluid flow in this region should be very valuable.

6. Visual observation of the air flow

A ' direct shadow ' method was employed as shown in Fig. 3.

The light from a high-pressure mercury lamp A was focused by a condensing lens system L on to an adjustable diaphragm at C, the centre of curvature of the spherical concave mirror M of 8 ft. focal length and 12 in. diameter. The returned image of the

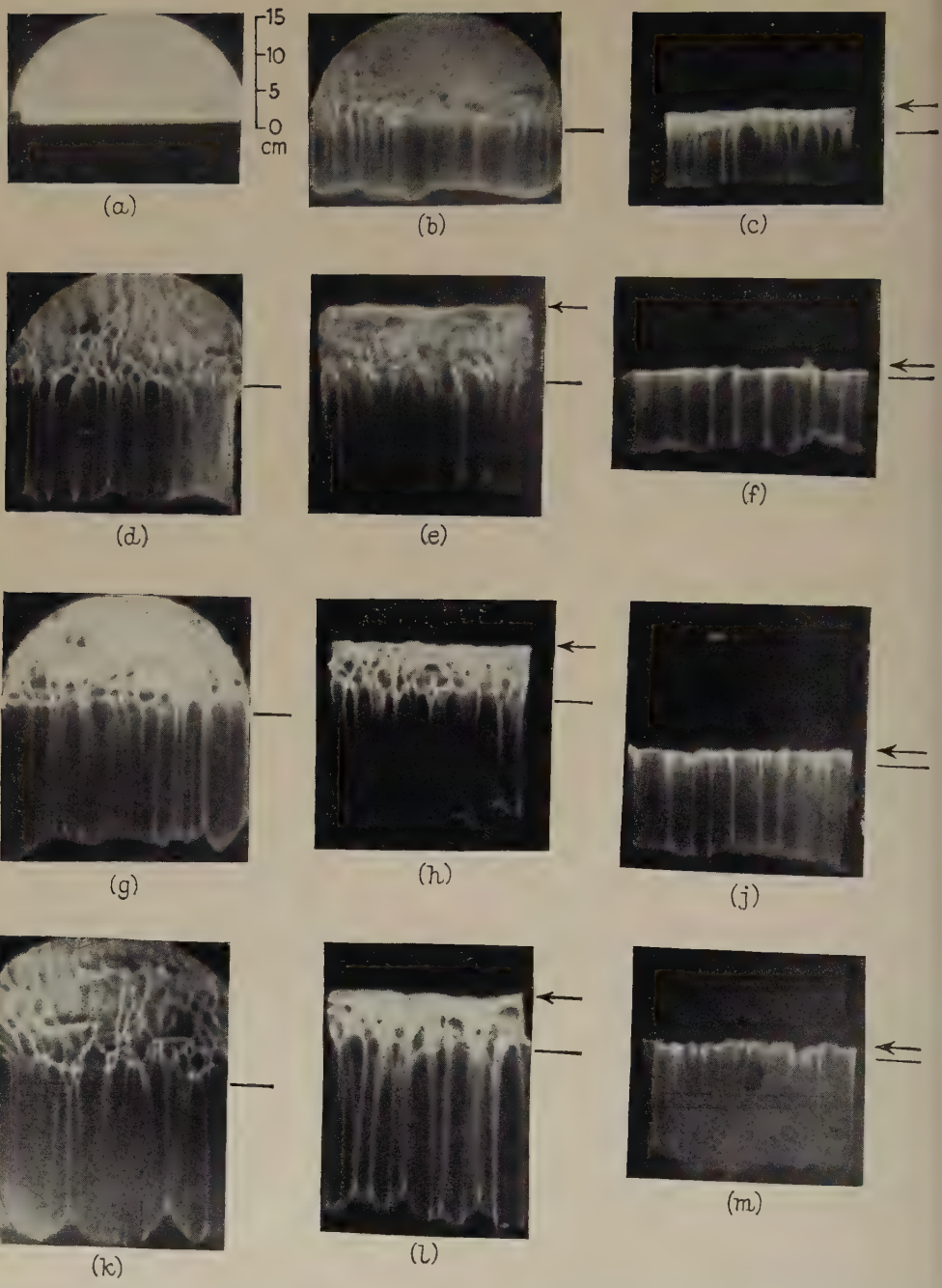

Figure 4. Patterns produced by the air above a heated horizontal plate. The position of the plate image in the absence of heating is indicated in each case by a line. (a) The unheated plate, with no upper cooled surface. The semi-circular outline is that of the mirror. (b) Plate temperature = 37°C. Open system (i.e. Ra = ∞). (c) Plate temperature = 37°C. Separation $d = 3$ cm. $Ra = 3{\cdot}6 \times 10^4$. (d) Plate temperature = 48°C. Open system. (e) Plate temperature = 48°C. Separation $d = 10$ cm. $Ra = 220 \times 10^4$. (f) Plate temperature = 48°C. Separation $d = 1{\cdot}3$ cm. $Ra = 0{\cdot}5 \times 10^4$. (g) Plate temperature = 60°C. Open system. (h) Plate temperature 60°C. Separation $d = 7$ cm. $Ra = 94 \times 10^4$. (j) Plate temperature 60°C. Separation $d = 2$ cm. $Ra = 2{\cdot}2 \times 10^4$. (k) Plate temperature = 70°C. Open system. (l) Plate temperature = 70°C. Separation $d = 7$ cm. $Ra = 114 \times 10^4$. (m) Plate temperature = 70°C. Separation $d = 1{\cdot}3$ cm. $Ra = 0{\cdot}7 \times 10^4$.

Except for Figs. (a), (b), (d), (g) and (k) an upper plate, maintained at 20°C was at the position indicated by the arrows.

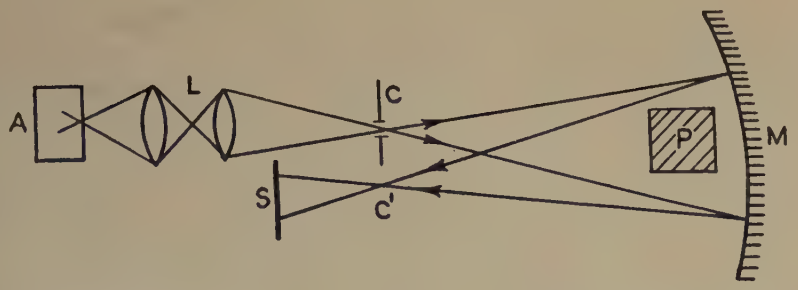

Figure 3. The diagram represents a plan view of the system, the heated surface *P* being at the same height as the pole of the mirror.

ntense point source at C was formed at C′, the light being very slightly displaced to allow an image to be formed on a screen S, placed at a suitable distance behind C′. As the working distance CM was 16 ft and the displacement CC′ was only some 4 in., very little aberration was introduced. The heated surface P was placed near the pole of the mirror, an inverted image being formed on the screen S. This method approximates very closely to that of a parallel-beam system in which the light passes twice through any phenomena above the plate P, the divergence in the beam being very small.

There are a number of interesting features in this arrangement, particularly convenient in that it is very readily converted into a conventional Schlieren system by the inclusion of a knife edge at C′. The light beam passes twice through the disturbed region, thus increasing the sensitivity. At the same time no significant doubling of the image occurs because of the almost exact return of the beam and the close proximity of the plate to the mirror.

The main advantage is the formation of a mirage-type image on the screen S of those phenomena close to the surface involving very marked changes in density gradient, which cause a vertical displacement of the light. This image is formed on the screen within the main image of the plate itself and hence is readily discernible against a dark background. The deviation of a ray passing through the phenomena is a measure of the spatial average temperature gradient over the path it has traversed, and hence the final displacement of the light on the viewing screen S is determined by this average gradient.

Observations were made, for a range of surface temperature, of the conditions above the open surface, and in enclosed air layers by the inclusion of a cooled upper surface at varying heights. Some of these observations are shown in Fig. 4 (Plate VI). The vertical filaments of light which characterize these images show that marked density changes occur in the air very close to the surface. Their presence is consistent with the existence of fluid elements, distributed across the surface of the plate, within which the density decreases rapidly with height. The inclusion of suitable stops in the field of view showed that these elements extend to a height between 15 and 20 mm above the surface. It must be noted that these elements do not lie in a single plane, but that the pattern formed is the integrated effect along the total path of the beam.

The most striking characteristic of the flow patterns is the strong suggestion of a cellular structure in the fluid motion. Although in the open system the vertical fluid elements, and hence the corresponding light filaments, are subject to occasional horizontal movement, the main impression is that of a regular maintained system of elements with a characteristic separation between them.

Moreover observation of the movements of the light filaments shows that these are formed in pairs, so that a fluid motion must be postulated which will give rise to two vertical light filaments simultaneously. This immediately suggests the Bénard cell, in which cooled air will descend at the centre with warmer air ascending at the boundaries, the paired vertical filaments resulting from light passing through sections of opposite walls of the cell. The patterns obtained cannot give an unambiguous measure of the cell

width as those observed do not necessarily lie in the same vertical plane above the pla
However, the observed separation of the light filaments is such as to suggest that the orc
of magnitude will be from 3 to 4 cm. This suggests a cell structure of a width appro
mately twice the height, and this is the actual shape of the cell predicted by Rayleigh (191

Furthermore, some of the observations on enclosed air layers were made at su
values of the Rayleigh number as to ensure the presence of cellular convection. Th
flow pattern in these cases shows them to be very similar to those in the lower regions
the open system and the enclosed layer of high Rayleigh number (i.e., large plate sepa
tion). This would indicate that a form of cellular convection is present close to the heat
surfaces in all the cases examined. The value of the Rayleigh number for the layer co
taining the motion can be calculated, this being a layer bounded by one rigid and one fr
plane surface. Assuming the thickness d of the layer to be 2 cm. in all cases, a range
values from 15,000 to 2,000 is obtained covering the highest to the lowest temperatu
used. These values are consistent with the existence of this mode, the critical value f
the onset of which is, in this particular case, 1101. It is not exactly clear what the upp
limit for its presence may be, but a value of the same order of magnitude as that for tw
rigid surfaces, namely 45,000, might be expected.

Schlieren observations on the open system indicate that above the cellular regic
convection continues by rising plumes which appear to originate as the result of a mush
rooming effect from the upward flow of the cellular motion. Some of the horizontal fila
ments of light above the cellular region in the direct shadow images may arise from thes
plumes.

The schlieren system was also used to establish that there were no cross-draught
or other extraneous disturbances to the region above the plate. The arrangement adopte
had the merit that there were no enclosing walls in close proximity to the heated plate
This was shown to be important by a series of subsidiary experiments on systems c
widely-differing dimensions working over appropriate temperature ranges. These demon
strated the very ready establishment, in the presence of screening walls, of a large-scal
circulation with warm air flowing up the screening walls with a compensating down draugh
in the centre of the plate. These experiments included a roughly cubical enclosure simila
in size to that employed by Thomas and Townsend, who however assert that subsidiar
experiments showed that no such large-scale circulation was present during their measure
ments.

7. Conclusions

Both the visual observations and temperature measurements show the existence above the heated surface, in all cases examined, of a mode of convection similar to that of the cellular convection previously noted in shallow enclosed layers. This motion is contained in a layer some 2 cm in extent, both above the open surface and in enclosed layers with a wide separation. This would of course lead one to expect the observed similarity in the heat transfer in the two cases.

The measured temperature profiles cover this cellular region, and for values of Z/Z_0 greater than 2·2 the results may be expressed quite well by the non-dimensional relation

$$\theta_0^{-1} \ln \frac{T}{T_A} = 3{\cdot}1 \left(\frac{Z}{Z_0}\right)^{-\frac{1}{2}}$$

This measured distribution is comparable with the prediction of the Priestley similarity theory, though the conditions postulated by that theory must be expected to apply more precisely in the higher layers. It is intended subsequently to extend the range of measurement using a very large laboratory system.

Acknowledgment

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A numerical study of the topographical factor in lee-wave amplitud

By C. E. WALLINGTON
Meteorological Office, Dunstable

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SUMMARY

A numerical study of the effect of mountain shape on two-dimensional lee-wave flow shows that asy metrical ridges are slightly more favourable than symmetrical profiles for the formation of lee waves, especi for short wavelengths. A technique is suggested for determining the lee wavelength when only part of lee-wave flow is observable.

1. INTRODUCTION

In several studies of two-dimensional airflow across mountain ridges the lee-wave amplitu is obtained as product of two distinct factors, an airstream factor determined entirely by t upper winds and temperatures of the airstream and a mountain factor determined by the wavelength and shape of the mountain. (See, for example, Scorer 1949). For mathemati convenience the cross-sections of the infinitely-long hill ridges used in these studies are specifi by the equation

$$h = Hb^2/(b^2 + x^2) \quad . \quad . \quad . \quad . \quad (1$$

where h is the height at a horizontal distance x from the summit whose height is H, and b is width parameter. This type of profile, which in this paper will be referred to as an elementa hill ridge, forms a useful basis for studying lee-wave flow but its symmetry is somewhat unrealist Most hill ridges are steeper on one side than on the other and those with gentle windward slop leading to steep lee escarpments appear from observation to be particularly suitable generato of lee waves.

The aim of this investigation is to study the effect of mountain shape on lee waves by compu ing the amplitude factors and positions of lee waves associated with a series of asymmetric ridges and with a real mountain range. It should be noted that the method used applies only studies of two-dimensional inviscid flow in which the wave amplitude is small enough to l studied by perturbation techniques.

2. ANALYSIS AND SYNTHESIS OF MOUNTAIN PROFILES

Any topographical cross-section may be described as the sum of a number of elementa ridges of the type specified by Eq. (1), i.e. by the formula

$$h = \sum_{r=1}^{r=n} \frac{H_r b_r^2}{b_r^2 + (x - x_r)^2} \quad . \quad . \quad . \quad . \quad (2$$

where n is the number of elementary ridges required to represent the cross-section and x_r denote the horizontal distance of the summit of each ridge from an arbitrary origin.

Scorer's (1949) treatment of two-dimensional flow enables the streamline displacemen ζ in lee waves downstream of an elementary hill ridge to be computed as

$$\zeta = - Hb \exp(-2\pi b/W)\, \xi \sin(2\pi x/W) \quad . \quad . \quad . \quad (3$$

where W is the lee wavelength and ξ is a function of the airstream wind and temperature charac teristics only. (A numerical study of values of ξ for some real airstreams has been made b Wallington and Portnall 1958).

The displacement in lee waves due to the profiles of Eq. (2) is given by the equation

$$\zeta = - \sum_{r=1}^{r=n} H_r b_r \exp(-2\pi b_r/W)\, \xi \sin[2\pi (x - x_r)/W] \quad . \quad . \quad (4$$

which can be manipulated into the form

$$\zeta = - M \xi \sin[2\pi (x - X)/W] \quad . \quad . \quad . \quad . \quad . \quad (5$$

where

$$M^2 = C^2 + S^2$$

$$X = W \tan^{-1}(S/C)$$

$$C = \sum_{r=1}^{r=n} H_r b_r \exp(-2\pi b_r/W) \cos(2\pi x/W)$$

$$S = \sum_{r=1}^{r=n} H_r b_r \exp(-2\pi b_r/W) \sin(2\pi x/W)$$

Thus the effect of a mountain range on lee waves may be studied by analysing its cross-section into a number of elementary ridges and then computing the ' mountain factor,' M, and the ' phase displacement,' X, for various lee wavelengths.

In this investigation each mountain profile studied was specified by 80 values of its height at equal horizontal intervals along its cross-section. The maximum height and its abscissa were taken to be H_1 and x_1, and the associated value of b_1 was taken as half the distance between the two points nearest to, and either side of, the summit where the height equalled $\frac{1}{2}H$. Fig. 1 illustrates the procedure. The values H_1, b_1 and x_1 were then used as the height, width and position parameters of an elementary ridge in the cross-section and the 80 heights were reduced by appropriate values of $H_1 b_1^2/[b_1^2 + (x - x_1)^2]$. The maximum numerical value of these reduced heights was taken to be H_2 and the process of finding the associated position and width parameters and reducing the 80 heights was repeated. Thus the analysis of the mountain range into successively smaller ridges or valleys was continued until the parameters of 15 elementary ridges were obtained. The number 15 happed to be convenient for the high-speed computing method used and this number appeared to be large enough for representation of all but the smallest features of the cross-section. The residual features omitted by the analyses made in this study were of the order of 2 to 8 per cent of the height of the primary elementary ridge.

Once the values of H_r, b_r and x_r had been obtained the ' mountain factor,' M, and ' phase displacement,' X, were calculated for various values of the lee wave-length.

3. Asymmetrical ridges

An asymmetrical cross-section can be arbitrarily but conveniently described by Eq. (1) if b is given one value, say B, when $x > o$ and another value, B', when $x < o$. The shape of such a ridge is shown in Fig. 2. Mountain factors and phase displacements for several ridges of this type were computed (by the method described in Section 2) for values of W ranging from 2 to 45 km and the nature of the results is illustrated in Figs. 3 and 4. Fig. 3 shows the variation of the ' mountain factor ' with wavelength for the three ridges sketched at the top of the figure. These three ridges are all 1 km high and, with each of them having 0·5 km for their value of B, they differ only in their upstream shapes. Ridge (i) has a symmetrical profile; (ii) has $B' = 1{\cdot}5$ km and in (iii) a gentle windward slope arises from a value of B' equal to 4·5 km. At short wavelengths the ' mountain factors ' for these three ridges are somewhat similar but as the wavelength increases beyond 10 km the curves diverge with ridge (iii) becoming increasingly more effective in producing lee waves than the other two. That this effect is due mainly to the total hill width rather than shape is illustrated in Fig. 4 which shows similar graphs for ridge (iii) and two others of equal height and equal width; $B + B' = 5$ km for each of the ridges (iii), (iv) and (v). Nevertheless the asymmetrical ridges produce slightly larger ' mountain factors ' than the symmetrical profile, especially for short wavelengths.

The vertical speed in a lee-wave flow is equal to $U\,\partial\zeta/\partial x$ where U denotes wind speed in the undisturbed flow. From Eq. (5) we have

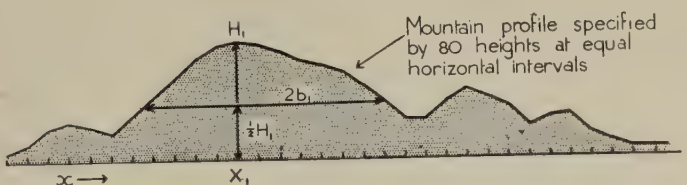

Figure 1. The mountain profile specified by 80 heights at equal horizontal intervals is analysed into ' elementary ' ridges of the type described in the text.

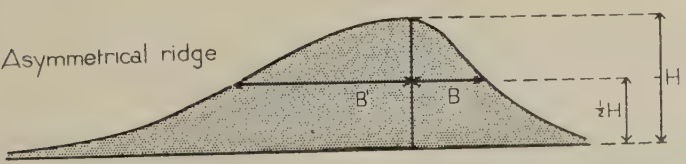

Figure 2. An asymmetrical ridge can be arbitrarily but conveniently specified by the upwind and downwind halves of the two different ' elementary ' ridges.

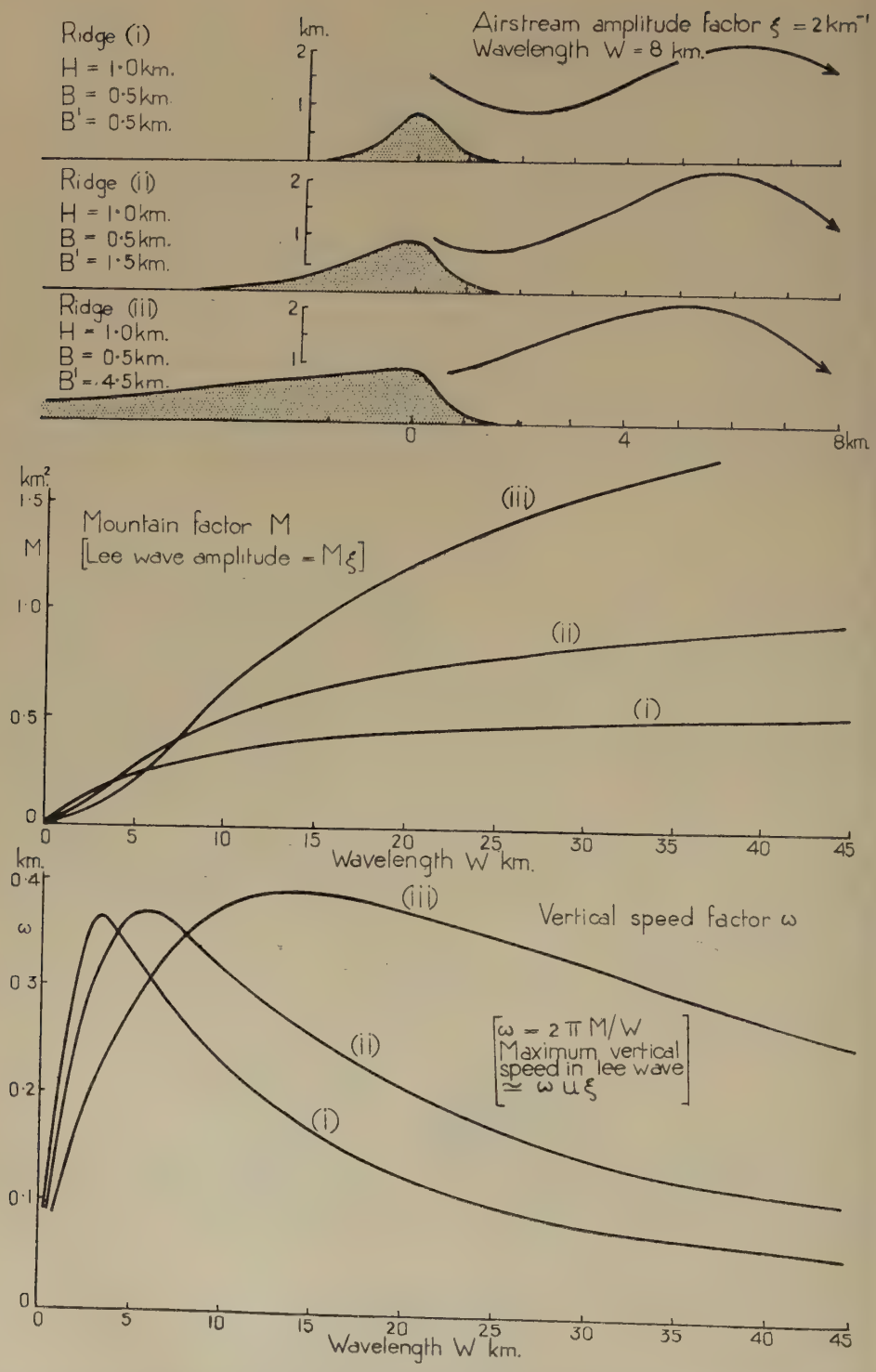

Figure 3. The gentler the upwind slope of an asymmetrical ridge the closer the first lee wave trough is to the summit of the ridge and the broader is the band of wavelengths for large ' mountain ' and ' vertical speed ' factors. Numbers on the plotted curves refer to the three ridges sketched.

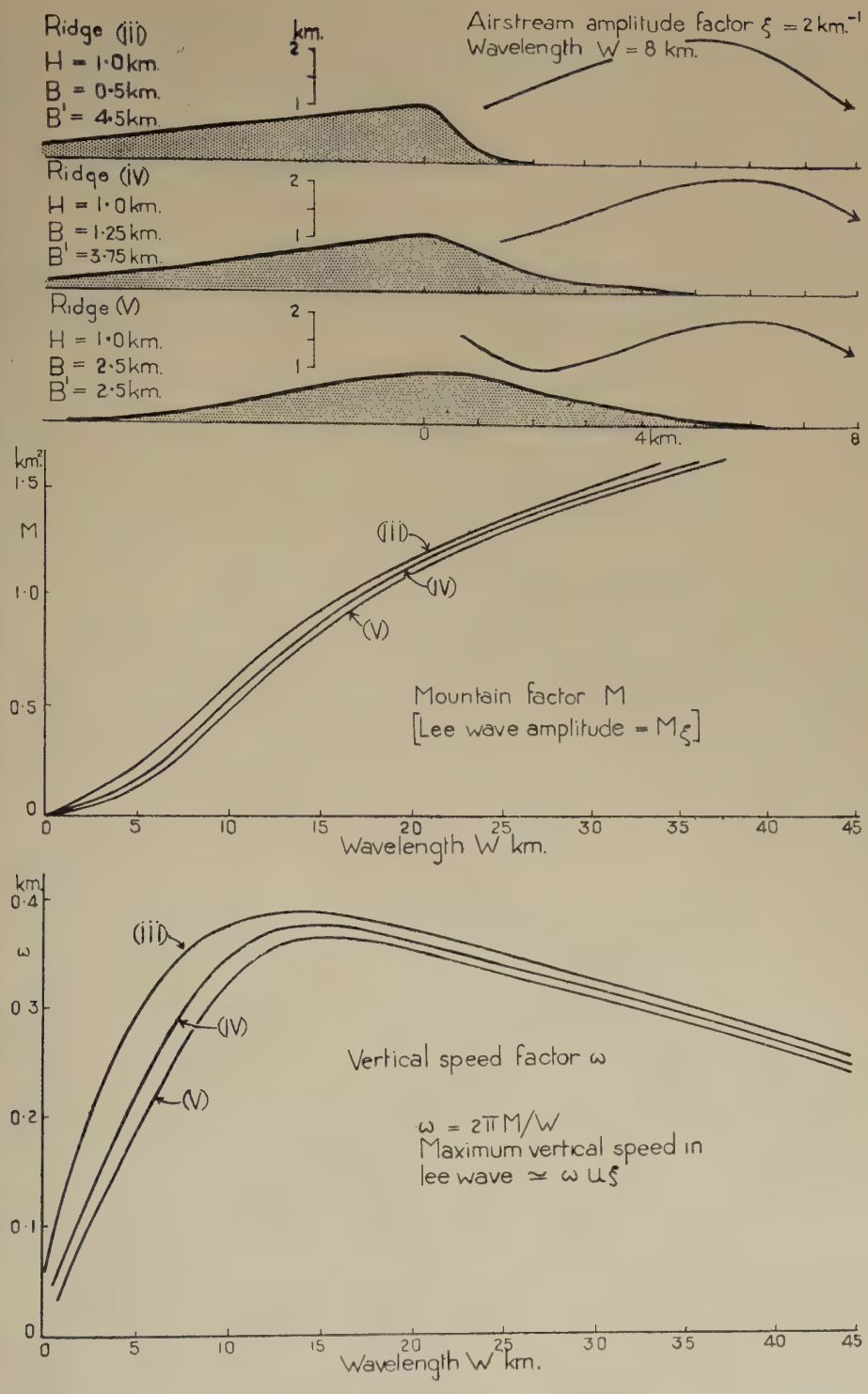

Figure 4. For mountain profiles of similar heights and overall widths the asymmetrical ridges are slightly better generators of large-amplitude waves than the symmetrical shape.

$$U\frac{\partial \zeta}{\partial x} = (2\pi M\, U\, \xi/W)\cos\,[2\pi\,(x - X)/W] \quad . \quad . \quad . \quad (6)$$

so that the maximum vertical speed is $wU\xi$, where $w = 2\pi M/W$ and may be termed a 'vertical speed factor' associated with the ridge. Values of this factor for a range of wavelengths applied to the five ridges are illustrated in the lowest sections of Figs. 3 and 4. The computed results show that although small symmetrical ridges are suitable for producing appreciable vertical speeds in short waves the asymmetrical profiles are more effective over a greater range of wavelengths, whether they are compared with small hills or with hills of equal total width but less asymmetry. These results were typical of those computed for ridges with other values of B and B'.

The asymmetry of a ridge also has an effect on the phase of the lee wave, and the nature of this effect is illustrated in Figs. 3 and 4 by the streamlines depicting the lee-wave part of the flow only. These streamlines have been calculated for a value of $\xi = 2$ km^{-1} at the 1·5 km level. Computations of ξ by Wallington and Portnall (1958) for 13 airstreams yielded a mean maximum of 0·86 km^{-1}, and in two cases just exceeded 2 km^{-1}. Thus the value of 2 km^{-1} for ξ is rather high but not exceptionally for real airstreams.

Study of these lee wave streamlines shows that by changing the hill profile from the symmetrical shape (i) to that of ridge (ii) the lee wave pattern is brought closer to the hill; it would therefore be reasonable to expect a strong flow of air down the lee slope of the ridge for this particular illustration. However, any further lengthening of the windward slope, as in ridge (iii), brings the wave so far forward that the streamline soon turns upwards into the first lee-wave crest; instead of augmenting a flow down the lee slope the lee wave now helps the flow to turn, or even break away from the mountainside, leaving relatively light winds at the foot of the escarpment. The lee wave due to ridge (iv) also favours a breakaway of the flow from the mountainside, but at a point further down the lee slope than that likely in the flow past ridge (iii). Over ridge (v) the lee-wave flow is conducive to a local zone of strong surface winds.

Although these features are described for one particular pair of values of ξ and lee wavelength, they do illustrate the nature of the effect of hill shape on the lee flow. In general a steepening of the lee escarpment or a lengthening of the windward slope brings the first lee trough closer to the mountain crest. Substantial asymmetry in this direction may lead to eddies in lee of steep escarpments while gentle lee slopes of large hills may have locally strong surface winds, but the precise details of the low-level flow is so intricately dependent on both topography and the airstream characteristics that there do not appear to be any simple criteria for the type of flow likely in any given situation.

4. A real mountain range

In field study of mountain airflow it is sometimes difficult to measure lee wavelengths because the distances between the principal mountain ridges in the terrain being explored are often too short for long wave trains to be established. For example, the thoroughly explored wave flow in lee of the Sierra Nevada, U.S.A., often has only one wave crest and two troughs over the Owens Valley (See Fig. 5). Since these troughs are usually close to the adjacent mountains the distance between them is not a reliable measure of the lee wavelength characteristic of the airstream.

It is, therefore, of interest to study the relationship between lee wavelength and wave position by means of the numerical method already described. Fig. 5 shows the results of the method applied to a 130 km cross-section of the high ground west of Independence, in the Owens Valley. The mountain factors and vertical speed factors have conspicuous maxima at lee wavelengths of about 11 km and 29 km. The computed positions of wave crests and troughs for lee wavelengths from 5 to 40 km show that the distance between the first lee-wave crest and the summit of the Sierra Nevada increases somewhat erratically with increasing lee wavelength.

A guide to the reliability of these results may be obtained by noting the positions of wave crests and troughs observed during the field study directed by Holmboe and Klieforth (1957). With preference being given to crests or troughs near the centre of Owens Valley rather than close to the mountainsides, these observed positions are inserted in Fig. 5 at the lee wavelengths where they appear to be most consistent with the computed positions of troughs and crests. It is encouraging to see that the lee wavelengths deduced in this way are consistent with the principal maxima and intervening minimum of the computed 'mountain' and 'vertical speed factors.'

The observation on 13 April 1955 appears at first sight to be incompatible with the computations. But this observation was made at a height of 13 km, and it seems very likely that the wave crest and trough were part of a wave system which included a nodal surface at some lower level. Most observed lee waves have amplitudes which attain a maximum in the lower troposphere and decrease to small values at high levels, but Wallington and Portnall (1958) have found that on occasions the wave flow contains a nodal surface above which the amplitude increases to its

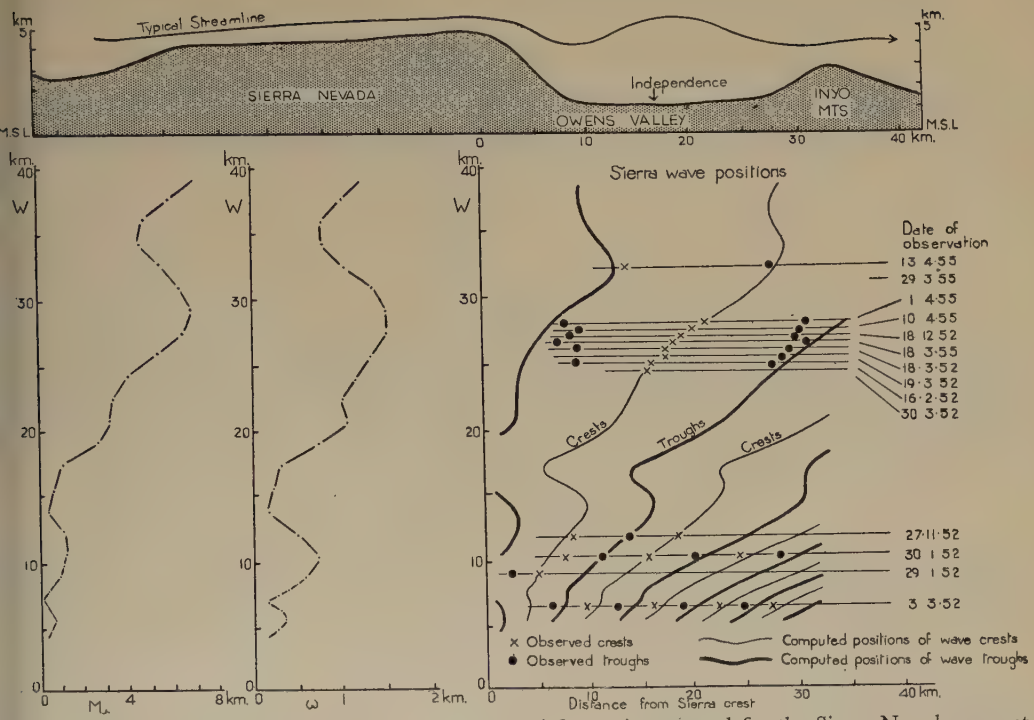

Figure 5. The ' mountain factors ' and ' vertical speed factors ' computed for the Sierra Nevada suggest that lee waves are likely to be more pronounced for two bands of lee wavelengths centred at 11 km and 29 km than for wavelengths of about 15 km. This conclusion is supported by wavelengths deduced from observations and computed wave positions. The observation used for 29.3.55 indicated the lee wavelength but not the position of the wave troughs and crest.

most significant maximum in the upper troposphere. This structure is associated more with long waves than with short wavelengths and it appears that the wave flow of 13 April 1955 was an example of this type of flow.

5. Conclusion

This numerical study shows that asymmetrical hill ridges are more effective than a symmetrical variety for setting off lee waves over a wide range of wavelengths.

If the lee wavelength appropriate to the airstream can be calculated, then the technique of the present paper enables the expected position of the troughs and ridges to be calculated. A comparison between observed and computed positions then provides a method of verifying the wavelength calculation even when the principal mountain ridges are too close together for long-wave trains to be established.

The computing method described in this study should be used with caution; it could give misleading results when the assumption of two-dimensional flow is not applicable to the air flow being studied. Errors could also arise because of the sensitivity of the computed ' mountain ' and ' vertical speed ' factors to changes of wavelength when the behaviour of very short waves over particularly rugged terrain is being studied. Nevertheless, when applied to wavelengths of between, say, 10 km and 40 km the method can be a useful complement to numerical techniques for studying the lee wavelength and amplitude characteristics of airstreams.

Acknowledgment

I am indebted to the Director-General of the Meteorological Office for permission to publish this paper.

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J. K. BANNON

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Stratospheric temperatures over the Antarctic

By J. K. BANNON
Meteorological Office, Harrow, Middlesex

(Manuscript received 22 May 1958)

SUMMARY

Some recent observations together with data from earlier expeditions (Court 1942, Schumacher 1955 and others) suggest that the annual variation of temperature in the stratosphere over the Antarctic is characteristically asymmetric with a steep rise of temperature in the spring. This is probably not general over the Arctic. The tropopause over the Antarctic is significantly higher than over the Arctic in their respective winters and the variability of temperature in the lower stratosphere appears to be considerably less in the same season over the Antarctic than over the Arctic.

1. INTRODUCTION

The main features of the distribution of temperature in the troposphere and lower stratosphere over the Antarctic have been described by Court (1942) and Schumacher (1955); the lower stratosphere is much colder in winter (difference of order 15°C) and a little warmer in summer (by 3°C or 4°C) than the stratosphere in the corresponding seasons over the Arctic. In the Antarctic winter, temperature continues to decrease with height above the check in the lapse rate at about 10-12 km which is identified as the tropopause, and the lapse rate in this layer above the tropopause is sometimes comparable with values in the troposphere. Court in fact suggested that the tropopause did not exist on many occasions in winter over Little America (about 78°S), but Schumacher says that there is usually a definite discontinuity in lapse rate about 10-12 km which, though it may not satisfy the conventional definition of tropopause as the lapse rate above it is more than 2°C per km, yet is a physically recognizable and significant level which should be called the tropopause. It is only on very rare occasions that the conventional tropopause cannot be found over the Arctic at usual levels (8-11 km).

2. RECENT OBSERVATIONS

In the last two years many observations have been made in the upper air over Antarctica. Unfortunately, although these observations are broadcast by radio, few of them have been published and scientists will probably have to await the publication of all the observations in microcard form by the I.G.Y. Meteorological Data Centre (World Meteorological Organization 1957).

However, the results of many soundings in the year June 1956 to May 1957 inclusive were published in the *New Zealand Daily Weather Bulletin*. The best series of observations in that year was for Williams Base (77½°S, 166°E) not far from Little America; only a few observations were missed and heights of at least 16 km (100 mb) were usually reached. Fig. 1 shows the monthly mean temperatures at 100 mb (16 km approximately) at Williams Base for the period June 1956 to May 1957 and also the averages for three other southern latitudes compiled from records for one or more stations for varying periods (Sources : Schumacher (1955), values quoted by Goldie, Moore and Austin (1958), Dewar (1958) and Meteorological Office (1957)). It is hoped that by combining observations from different longitudes, local effects will have been eliminated to some extent. Fig. 2 shows monthly mean temperatures for three levels at Halley Bay (75½°S, 26½°W) since April 1957, as broadcast by radio.

3. ANNUAL VARIATION OF TEMPERATURE

It is evident from Figs. 1 and 2 that there is a steep rise in temperature in the stratosphere in spring. Court (1942) and Schumacher (1955) noted this also for Little America and Maudheim respectively and it would appear that this is general over the Antarctic and that the general shape of the annual variation of temperature is similar at all latitudes. This skewness of the annual variation of temperature in the stratosphere is observed at a few Arctic stations but not at all at others (Goldie, Moore and Austin 1958). It is difficult to generalize for the Arctic because

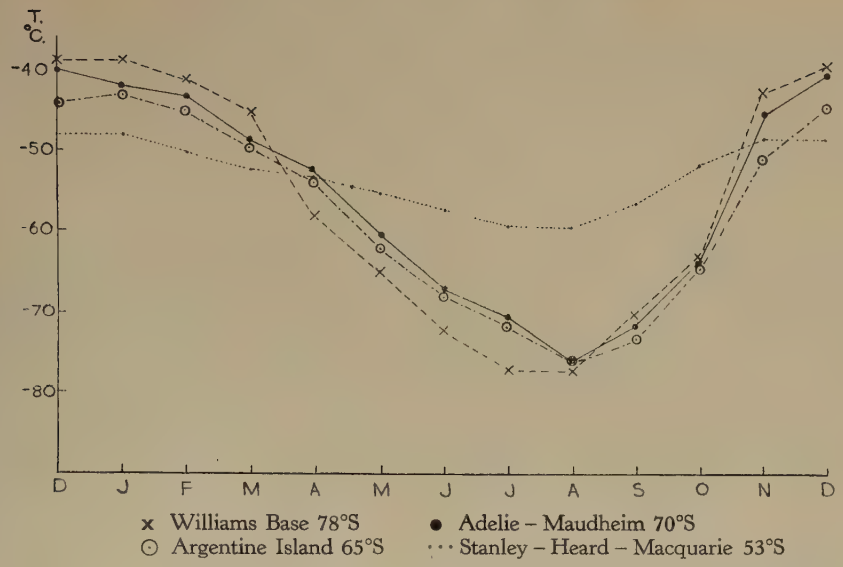

Figure 1. Average monthly temperatures at 100 mb (16 km) for four southern latitudes. Periods: Williams Base 1 year; Adelie Land 1 year; Maudheim 1 year 9 months; Argentine Island 3 years; Heard Island 3 years; Macquarie Island 3 years; Stanley 5 years.

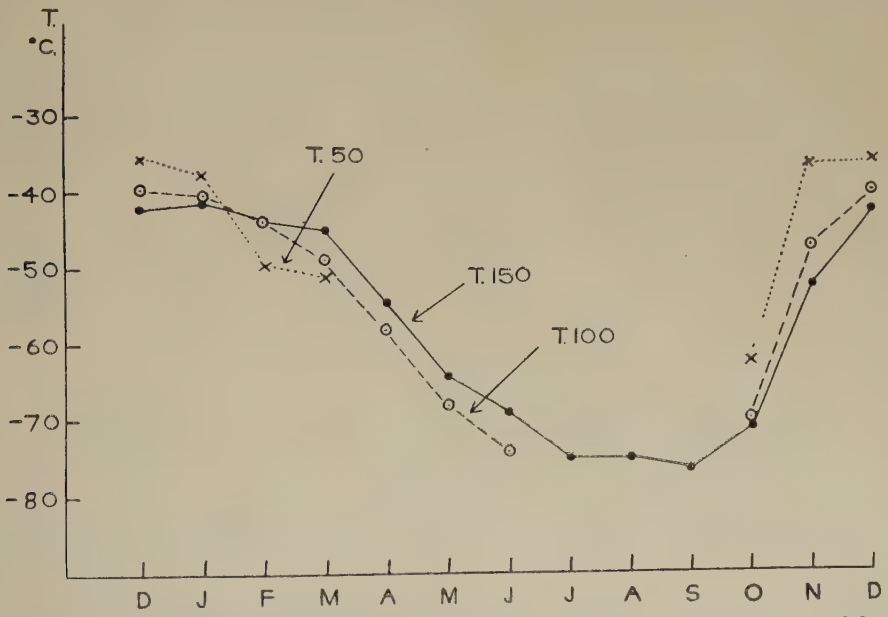

Figure 2. Average monthly temperatures in the stratosphere, Halley Bay, April 1957 to March 1958. T_{50} temperature at 50 mb (20 km); T_{100} temperature at 100 mb (16 km); T_{150} temperature at 150 mb (13 km).

observations are not available for a large section (U.S.S.R.) and because the general upper-air pattern is not at all symmetric about the pole, but it would appear that this asymmetry of the annual variation of temperature in the lower stratosphere over the Antarctic is not a general feature of the Arctic.

4. The tropopause

The soundings from Williams Base confirm earlier observations that the tropopause as conventionally defined often disappears in midwinter. In August 1956 there was no conventional tropopause below 14 km on half of the number of soundings. The fall of temperature with height was not large, however, in the layers above about 10 km.

The average height of the tropopause above Maudheim (Schumacher 1955) in midwint was in the region of 11 km (200 mb); in July and August 1956 it was about 10 km (240 m over Williams Base. Average heights of the tropopause over the Arctic in midwinter are signi cantly lower, about 9 km (300 mb) (Moore 1956). Dewar (1958) has found that the tropopau above Argentine Island (65°S, 64°W) in winter is appreciably higher than at similar latitudes the northern hemisphere at the same season.

5. Variability of temperature

In winter, temperatures in the lower stratosphere over the Arctic may be low (about − 85°C but can also be almost as high as in summer (about − 45°C). The two types of stratosphere we first pointed out by Riehl (1950). Recently Austin and Krawitz (1956) have suggested that larg scale dynamical developments take place in the winter stratosphere with considerable vertic motion and this seems the most reasonable explanation of these large variations in temperatur

TABLE 1. Mean temperature (°C) and range over a month at about 19 km

	Williams Base 77½°S (1 yr)	Argentine Is. 65°S (3 yrs)	Alert 82°N (3 yrs)	Barrow 71°N (5 yrs)	Lerwick 60°N (5 yrs)
Midwinter (July or January)					
Mean	− 82	− 79	− 65	− 51	− 62
Range	11	*c* 23	39	32	25
Spring (October or April)					
Mean	− 60	− 61	− 50	− 51	− 52
Range	27	40	20	14	30

The observations from Williams Base showed that the variation of temperature in the strato sphere was not large in midwinter. Table 1 gives mean temperatures and the range of temperatur over a month in winter and spring at about 19 km for Williams Base, Argentine Island and also for comparison, for Alert, Barrow and Lerwick. The 16 observations made at Little Americ (78°S, 165°W) between 1 July and 28 August 1940, which reached 14 km or thereabouts, showed variation of temperature at that level of only 10°C; observations for higher levels were seldo available (Court 1949). Schumacher (1955) makes no mention of variability of temperature in th stratosphere over Maudheim. It would appear that the large variations noted over the Arcti do not occur over the Antarctic although more data are required to confirm this suggestion, whic implies an interesting difference between dynamical developments in the stratosphere over th Arctic and Antarctic.

Stratospheric temperatures in the summer have quite a small variation (10°C or less) ove both Arctic and Antarctic.

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551.501.81 : 621.396.9

Radar echoes from atmospheric inhomogeneities

By R. F. JONES
Meteorological Office, London

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Summary

The possibility that radar echoes from atmospheric inhomogeneities are the explanation of echoes received in conditions which make it certain they are not received from precipitation elements ('angel' echoes) is examined quantitatively. It is shown that such an explanation is theoretically possible but that the required refractive index changes are large, involving changes of a few mb in vapour pressure and/or a few degrees C in temperature in a distance of about 25 cm. Reflection of the main or side-lobe energy at a limited quasi-horizontal surface is shown to give the most likely meteorological explanation, but the inability to measure changes of refractive index in the free atmosphere to the required accuracy prevents positive confirmation.

1. Introduction

In a recent paper Harper (1957) gives examples of radar echoes received at East Hill using 10 cm wavelength equipment in weather conditions which make it certain that the echoes were not received from raindrops or cloud formations. The name 'angels' has been given to echoes of a similar character observed in America and has also been applied to unexplained echoes seen in large numbers on vertically-looking radar using wavelengths as low as 0·86 cm. Plank (1956) has given an exhaustive account of 'angel' activity and the theories to explain it, while other accounts have been given by Goldstein (1951), Gould (1952) and Jones (1952) among others.

The fact that echoes can be received from birds cannot be gainsaid but Harper (1957) reaches the conclusion 'that all widespread displays of 'angels' on centimetric radar in southern England are caused by birds.' Since many writers (e.g. Gordon 1949, Friend 1949), including the author of this paper (1952), have mentioned the possibility that the echoes may also be explained by scattering or reflection from atmospheric inhomogeneities, this possibility is here subjected to quantitative examination.

2. The observations

The observations here considered are those of the type reported by Harper and the author, i.e., seen on PPI and RHI radars. The echoes usually took the form, when seen on the PPI display, of a number of small dots, more or less evenly distributed around the station and similar to those received from small aircraft. The RHI display showed the echoes to be confined usually to a horizontal belt with the top on most occasions 5,000 ft or below. The vertical spread of an individual echo was no greater than would be expected of an echo from a point source. The angle of incidence of the radar beam did not appear to be of importance since the echoes were visible at short ranges with the beam elevation 15° to 19° as well as at ranges beyond 50 km when the beam was nearly horizontal. The echoes were observed under different weather conditions ranging from completely cloudless to mainly cloudy, and sometimes followed or preceded normal weather echoes. There was a marked tendency, if the echoes were observed during the afternoon, for the number of echoes to increase considerably in the period around sunset and thereafter remain numerous. Echoes moved from the general direction of the wind with speeds usually in excess of the wind speed, often about twice the speed of the probable wind.

3. Back-scattering from an atmospheric inhomogeneity

The theory of back-scattering from an atmospheric inhomogeneity has been dealt with by Goldstein (1951). Using this theory it can be shown (Jones 1952) that if an 'angel' echo is visible with the East Hill equipment at a range of 40 km or more and the echo arises by back-scattering

from an inhomogeneity, the gradient $\frac{d\mu}{dR}$ of refractive index along the normal to the scattering surface must be given by :

$$\frac{d\mu}{dR} = \pm 2{\cdot}51 \times 10^{-4}\ \mathrm{m}^{-1} \quad . \quad . \quad . \quad . \quad (1)$$

Furthermore this gradient must persist over an area normal to the beam of 10^4 m^2 over a distance $\delta R = 2{\cdot}5$ cm (one quarter wavelength) or more.

At a temperature of 273°K, pressure 900 mb and vapour pressure 5 mb the changes of vapour pressure (δe) and temperature (δT) which must occur over 2·5 cm to satisfy Eq. (1) are shown in Table 1.

TABLE 1. Changes in vapour pressure (δe) and temperature (δT) which must occur simultaneously over a distance of 2·5 cm to satisfy Eq. (1)

δe (mb)	± 3·0	± 2·5	± 1·5	± 1·0	± 0·5	0·0	∓ 0·5
δT (°C)	± 7·8	± 5·6	± 1·2	∓ 1·1	∓ 3·3	∓ 5·5	∓ 7·7

The changes are large, but not impossibly so, but it is difficult to imagine them persisting for any length of time over such an area normal to the beam (which it must be remembered in RHI applications sweeps from 0° to 19° in elevation). In addition direct back-scattering cannot explain echoes which move substantially faster than the wind speed. Direct back-scattering, therefore, is not now regarded as a possible explanation of a large number of ' angel ' echoes seen with the East Hill equipment, although it may well be the explanation of echoes seen using vertically-looking radars.

4. Reflection at a horizontal layer

In seeking to explain the occurrence of the echoes of the type observed by reflection at a horizontal layer we are saying that the echoes are received from *ground-based* objects with propagation of the radio energy to and from the target via a reflecting layer in the atmosphere. A restriction on the dimensions of the reflecting layer, however, is imposed by the observation that the echoes move. If there was a continuous reflecting layer throughout the area of observation then echoes from numerous ground objects would be received, but, with no change in ray path during the revolution of the transmitting and receiving aerial, the echoes would be stationary. The fact that the echoes are very small and move in the direction of the wind implies that the reflecting surface is of limited horizontal extent and the echoes are signals from different ground-based objects received via a moving reflecting layer. The limitations on horizontal dimensions are that the dimensions at right angles to the radar beam should be less than the beam-width and along the beam about half the pulse-length (300 m).

If horizontally polarized waves are incident obliquely on a layer in which the refractive index falls smoothly from μ to $\mu - \delta\mu$ in a small height interval z and ϕ_0 and ϕ are the glancing angles of incidence and refraction respectively we have, if $\frac{2\pi z}{\lambda}\sin\phi_0$ is small compared with unity (Appleton 1946)

$$\mu\cos\phi_0 = (\mu - \delta\mu)\cos\phi \quad . \quad . \quad . \quad . \quad (2)$$

and the reflection coefficient ρ giving the ratio of the *amplitudes* of the reflected and incident vibrations is

$$\rho = \frac{\sin(\phi_0 - \phi)}{\sin(\phi_0 + \phi)} \simeq \frac{\phi_0 - \phi}{\phi_0 + \phi} \quad . \quad . \quad . \quad . \quad (3)$$

for small values of ϕ, ϕ_0. $\left(\text{For vertical polarization also } \rho \simeq \frac{\phi_0 - \phi}{\phi_0 + \phi}\right)$.

In the atmosphere μ does not depart from unity by more than a very small amount, of the order of 10^{-4}, and $\delta\mu$ will be of the order of 10^{-5}, so that with negligible error we can write Eq. (2) in the form

$$\cos\phi_0 = (1 - \delta\mu)\cos\phi \quad . \quad . \quad . \quad . \quad (4)$$

from which it can be shown that when ϕ, ϕ_0 are small

$$\phi_0 - \phi = \frac{\delta\mu}{\phi_0}$$

when the glancing angle of incidence departs appreciably from the critical angle given by $\phi_0 = (2\delta\mu)^{\frac{1}{2}}$ from Eq. (4). For $\delta\mu$ of the order of 10^{-5} this critical angle is about 0·26°.

Hence for glancing angles of incidence greater than about 0·26° we have, using Eq. (3)

$$\rho = \frac{\delta\mu}{2\phi_0^2} \quad . \quad . \quad . \quad . \quad . \quad . \quad (5)$$

If we now consider a square horizontal layer of the type discussed of side d, height h above the horizontal plane through the transmitter and distance r (large compared with d and h) from the transmitter, the glancing angle of incidence ϕ_0 of the transmitted radiation falling on this layer will be given by $\sin\phi_0 \simeq \phi_0 = \frac{h}{r}$ and the angle in the vertical plane subtended by the distance d at the transmitter will be

$$\frac{d\sin\phi_0}{r} \simeq \frac{d\phi_0^2}{h}$$

If θ is the vertical beam-width of the transmitted radiation, supposed uniformly distributed through this beam-width, the proportion of the vertical beam-width intercepted by the layer will be $\frac{d\phi_0^2}{h\theta}$ for $\frac{d\phi_0^2}{h} < \theta$ and unity for $\frac{d\phi_0^2}{h} \geqslant \theta$.

If ψ is the horizontal beam-width the proportion of this dimension intercepted by the layer will be $\frac{d}{r\psi} = \frac{d\phi_0}{h\psi}$ for $\frac{d\phi_0}{h} < \psi$ and unity for $\frac{d\phi_0}{h} \geqslant \psi$.

If P is the power output of the transmitter the power P_i incident on the layer will be :

Case (i) $\quad \frac{d\phi_0^2}{h} < \theta, \frac{d\phi_0}{h} < \psi, \quad P_i = \frac{Pd^2\phi_0^3}{h^2\psi\theta} \quad . \quad . \quad . \quad . \quad (6)$

Case (ii) $\quad \frac{d\phi_0^2}{h} \geqslant \theta, \frac{d\phi_0}{h} < \psi, \quad P_i = \frac{Pd\phi_0}{h\psi} \quad . \quad . \quad . \quad . \quad (7)$

Case (iii) $\quad \frac{d\phi_0^2}{h} \geqslant \theta, \frac{d\phi_0}{h} \geqslant \psi, \quad P_i = P \quad . \quad . \quad . \quad . \quad . \quad (8)$

Case (iv) $\quad \frac{d\phi_0^2}{h} < \theta, \frac{d\phi_0}{h} \geqslant \psi, \quad P_i = \frac{Pd\phi_0^2}{h\theta} \quad . \quad . \quad . \quad . \quad (9)$

The *intensity* of the reflected radiation is proportional to the square of the amplitude of the vibrations so that the power reflected to the ground will be $\rho^2 P_i = \frac{(\delta\mu)^2}{4\phi_0^4} \cdot P_i$.

The power re-radiated from the ground is more difficult to assess. There are likely to be many obstructions which are normal to the incident radiation, since the glancing angle of incidence will be small, and which will therefore reflect back along the same track the energy incident upon them. It seems likely that, if anything, we shall underestimate the energy scattered back along the same track as the incident energy if we assume that the earth acts as a diffuse

surface scattering the incident energy uniformly in all directions. On this assumption it can be shown that the total power (P_r) received back at the receiver is given by :

$$P_r = \frac{A}{64\pi h^2} \frac{(\delta\mu)^4}{\phi_0^5} P_i \quad . \quad . \quad . \quad . \quad . \quad (1)$$

for small values of ϕ_0, where A is the area of the aerial aperture.

For the height-range equipment used at East Hill we have $P = 5 \times 10^5$ w, $\theta = \pi/12$, $\psi = \pi/24$ and $A \simeq 6$ m^2 and the half pulse-length, which is the limiting value of d, is 300 m.

Substituting these values in Eqs. (6) to (9) and applying Eq. (10) leads to :

Case (i) $$P_r = 3{\cdot}92 \times 10^{11} \frac{(\delta\mu)^4}{h^4 \phi_0^2} \text{ w} \quad \begin{matrix} \phi_0^2 < 8{\cdot}72 \times 10^{-5} h \\ \phi_0 < 4{\cdot}36 \times 10^{-4} h \end{matrix} \quad . \quad . \quad . \quad (11)$$

Case (ii) $$P_r = 3{\cdot}41 \times 10^{7} \frac{(\delta\mu)^4}{h^3 \phi_0^4} \text{ w} \quad \begin{matrix} \phi_0^2 \geqslant 8{\cdot}72 \times 10^{-5} h \\ \phi_0 < 4{\cdot}36 \times 10^{-4} h \end{matrix} \quad . \quad . \quad . \quad (12)$$

Case (iii) $$P_r = 1{\cdot}49 \times 10^{4} \frac{(\delta\mu)^4}{h^2 \phi_0^5} \text{ w} \quad \begin{matrix} \phi_0^2 \geqslant 8{\cdot}72 \times 10^{-5} h \\ \phi_0 \geqslant 4{\cdot}36 \times 10^{-4} h \end{matrix} \quad . \quad . \quad . \quad (13)$$

Case (iv) $$P_r = 1{\cdot}71 \times 10^{8} \frac{(\delta\mu)^4}{h^3 \phi_0^3} \text{ w} \quad \begin{matrix} \phi_0^2 < 8{\cdot}72 \times 10^{-5} h \\ \phi_0 \geqslant 4{\cdot}36 \times 10^{-4} h \end{matrix} \quad . \quad . \quad . \quad (14)$$

where h is measured in metres. These equations are applicable for values of ϕ_0 greater than the critical angle $(2\delta\mu)^{\frac{1}{2}}$.

Airborne refractometer measurements in America (Crain 1953) have given maximum values of $\delta\mu$ *occurring in a few hundred feet* of about 5×10^{-5}, but for the small vertical distances implied by the restriction $\frac{2\pi}{\lambda} z \sin\phi_0 \ll 1$ it is unlikely that values of $\delta\mu$ as large as this occurred. For conditions likely to occur in England in March (when many such echoes were observed) it may readily be shown that over small intervals of height, for which the pressure change may be neglected, $\delta\mu$ is given approximately by $10^6\,\delta\mu = 5{\cdot}1\,\delta e - 1{\cdot}2\,\delta T$. Since the change of vapour pressure δe cannot exceed the saturation-vapour pressure at the temperature assumed (273°K+ the maximum possible value of δe is about 6 mb, requiring an inversion of temperature of about 16°C to give a value of $\delta\mu$ of 5×10^{-5}. On the other hand a value of 2×10^{-5} would not be impossible. Such a value corresponds to a critical angle of $6{\cdot}3 \times 10^{-3}$ rad = 0·4°.

The minimum signal detectable by the East Hill equipment is about 10^{-13} w and in order to explain an echo observed at a height of 1,000 m (a frequent occurrence) by this mechanism the height, h, of the reflecting layer envisaged must be 500 m. Substituting $\delta\mu = 2 \times 10^{-5}$, $P_r = 10^{-13}$ w and $h = 500$ m in Eqs. (11) to (14) we find that there is no value greater than the critical angle which will satisfy them.

For smaller values of h reflection at higher angles of incidence would be possible but the displayed height ($= 2h$) would be correspondingly smaller, e.g. for $h = 100$ m, detectable signals would be received for angles of incidence up to 3·7°.

It appears therefore impossible to explain echoes appearing at heights of 1,000 m or more at angles of elevation greater than 0·4° by a straightforward reflection mechanism.

To explain these echoes we must postulate *some* energy transmitted at very low angles of elevation whatever the angle of elevation of the main lobe. That the radiation pattern of any transmitting aerial is not perfect is well known and some small part of the transmitted energy is dissipated in side-lobes. This energy is, however, a small fraction of the energy transmitted in the main lobe and it is clear that, since the main lobe of energy can only marginally give a detectable signal at angles greater than the critical angle, the energy transmitted in side-lobes could only possibly give a detectable signal if it were transmitted at angles less than the critical angle and suffered total reflection. We consider now, therefore, the received signal strength P_r for angles of glancing incidence below the critical value for which the reflection coefficient ρ is unity. This can be immediately obtained by multiplying the right-hand side of Eqs. (11) to (14) by $(2\phi_0^2/\delta\mu)^4$, i.e. $1/\rho^4$, with the additional restriction on ϕ_0 of $\phi_0 < (2\delta\mu)^{\frac{1}{2}}$.

Taking again a maximum value of $\delta\mu$ of 2×10^{-5} the over-riding restriction is $\phi_0 < 6{\cdot}3 \times 10^{-3}$ rad.

Thus, for all practical values of h, Case (i) holds with $\phi_0 < 6{\cdot}3 \times 10^{-3}$ except for values of h up to $63/4{\cdot}36 = 14{\cdot}5$ m when Case (iv) holds. Table 2 shows the received power for various values of h assuming $\phi_0 = 6{\cdot}3 \times 10^{-3}$; the Table also shows the range ($= h/\phi_0$) at which the reflecting surface must lie.

TABLE 2. RECEIVED POWER (P_r) FOR CRITICAL GLANCING ANGLE OF INCIDENCE OF $6{\cdot}3 \times 10^{-3}$ RAD (0·4°) AND VARIOUS HEIGHTS (h) OF LIMITED REFLECTING AREA

h (m)	P_r (w)	Range (km)
10	$2{\cdot}77 \times 10^{-5}$	1·6
20	$2{\cdot}5 \times 10^{-6}$	3·2
30	$4{\cdot}95 \times 10^{-7}$	4·8
100	$4{\cdot}01 \times 10^{-10}$	15·9
500	$6{\cdot}4 \times 10^{-12}$	79·5
1000	$4{\cdot}01 \times 10^{-14}$	159

With a minimum detecting capability of the receiver of about 10^{-13} w it is clear that easily detectable signals can be received from the main lobe of radiation by reflection at a limited area at considerable ranges. At short ranges and low heights of the reflecting area a reduction of transmitted power P by a factor 10^5 or greater would still give a detectable signal via reflection areas at ranges up to 5 km. Transmitted radiation patterns have not been measured to the required accuracy but it seems quite possible that such a small proportion of the transmitted radiation might be escaping along the elevation of 0·4° whatever the angle of elevation of the main lobe. The resulting echo, however, would be *displayed* as if it were received from the main lobe so that an echo received via a side-lobe and a reflecting surface 30 m high at a range of 4·8 km would appear on the display if the elevation angle of the main lobe was 10°, say, at a slant range of 9·6 km and apparent height of 9,600 sin 10° = 1,670 m.

When the elevation angle was 19° – the maximum angle of sweep of this height-range equipment – the display height would be 3,120 m at the same slant range. It is, therefore, possible to display echoes at heights from 3,120 m to the ground at a range of 9·6 km by the reflection of side-lobe energy at a surface only 30 m high and 4·8 km away. Variations in the side-lobe energy with angle of elevation would lead to a broken, rather than continuous, echo. Such echoes are often observed from aircraft at short ranges.

As this reflecting surface moves away from the transmitter the glancing angle of incidence ϕ_0 will decrease and the received power will decrease as ϕ_0^6 requiring a greater amount of side-lobe energy to give a detectable signal. This greater amount is likely only to be found in side-lobes close to the main lobe, so that echoes received by this mechanism at high angles of elevation of the main lobe are only likely to be received at short ranges, those received at longer ranges being detectable only on low angles of the main beam. This is a feature of the echoes so far observed.

If the side-lobe energy was 30 db below the energy of the main lobe (i.e., a factor of 10^{-3}) a signal via a reflecting surface 30 m high using this side-lobe would still be detectable when the surface subtended an angle at the transmitter of about $\frac{1}{3}$ the critical angle, i.e., at a range of 14·4 km, even though the power received depends on ϕ_0^6. This would be displayed at a range of 28·8 km and the height displayed would depend on the side-lobe used. The main lobe could contribute a signal at angles down to about $\frac{1}{10}$ of the critical angle, i.e. at a range of 48 km but at such ranges for such a low height the curvature of the earth must be taken into account and the reflecting surface would be below the radio horizon at a range of about 22 km. The main lobe would therefore give a signal of displayed height 60 m at ranges of 44 km in to a range of 0·5 km corresponding to an angle of elevation of 7° determined by Eq. (13).

It seems, therefore, that reflecting surfaces only 30 m above the ground could be responsible for echoes displayed on a height-range equipment at ranges out to about 44 km at a height of 60 m and at heights up to 3,000 m at ranges from 9·6 km to about 28·8 km. Similar results would

apply to other heights of reflecting surface but it must be remembered that for an echo to b received by the side-lobe mechanism, and therefore displayed at incorrect height, the reflectin layer must not subtend an angle at the transmitter greater than about 0·4°, i.e., it must not b nearer than the appropriate range shown in Table 2 with the displayed range double this valu Thus echoes displayed at short ranges (e.g., 10 km or less) and considerable heights (e.g., 1,000 r or more) can only be explained meteorologically by this mechanism if the reflecting surface is c the order of 30 m high. Higher layers can, of course, contribute to echoes at greater ranges.

Since different layers, if they exist, may contribute to give echoes, and the speed of movemen of the echo is always twice that of the layer, some variation in speed and direction of movemen between individual echoes is to be expected. The restriction on vertical distance z, over which the refractive index change $\delta\mu$ takes place, that $\frac{2\pi}{\lambda} z \sin\phi_0$ should be small compared with unity implies, for $\phi_0 = 6{\cdot}3 \times 10^{-3}$ rad, $\lambda = 10$ cm and regarding 10^{-1} as small compared with unity that z should not exceed about 25 cm.

7. Discussion and conclusions

Whether variations of 2×10^{-5} in refractive index can occur in such small intervals of height as 25 cm is at present unknown but it seems that if there is a *meteorological* explanation of these echoes, variations of this order must occur. The noticeable increase in number of echoes in the evening as the ground cools suggests that a tendency to stratification of the atmosphere near the ground may well be a contributory factor.

The magnitude of vapour-pressure change required – a few mb – makes it unlikely that the required gradients could ever occur at temperatures below about 0°C, where the saturation-vapour pressure is only about 6 mb, unless there was a very considerable inversion of temperature in a vertical distance of 25 cm. Echoes of the type discussed have not been observed at temperatures below 0°C.

The restriction that $\frac{2\pi}{\lambda} z \sin\phi_0$ should be much less than unity means that the appropriate change in refractive index $\delta\mu$ would be more easily attainable on long wavelengths than on short. This appears to be confirmed by the large number of echoes reported on wavelengths as high as 100 m or more, including echoes from the tropopause region which certainly could not be explained as echoes from birds. (See summary in Plank (1956)).

Even with narrow beam-width equipment, and the main-lobe contributing, the energy returned to the receiver is only a very small fraction of that transmitted; thus echoes of this type would only be expected to be observed with radars having high power outputs and narrow beam widths and this is confirmed by the observations so far reported.

Acknowledgments

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551.556.8

A note on empirical wind-wave formulae

By H. CHARNOCK
National Institute of Oceanography, Wormley, Surrey

(Manuscript received 28 May 1958)

1. Introduction

The aim of most empirical wind-wave formulae is to relate some statistic of the wave field with some statistic of the wind field and with an appropriately defined fetch. We shall only consider the waveheight, for which various measures are used such as the significant waveheight (the mean height of the highest one-third of the waves) and the equivalent waveheight (the height of a simple sinusoidal wave which would have the same energy as those observed). If the water surface can be described as a stationary random gaussian surface with a narrow spectrum all such statistics are simply related to the root-mean-square fluctuation of the level of the water surface at a point (Longuet-Higgins 1952).

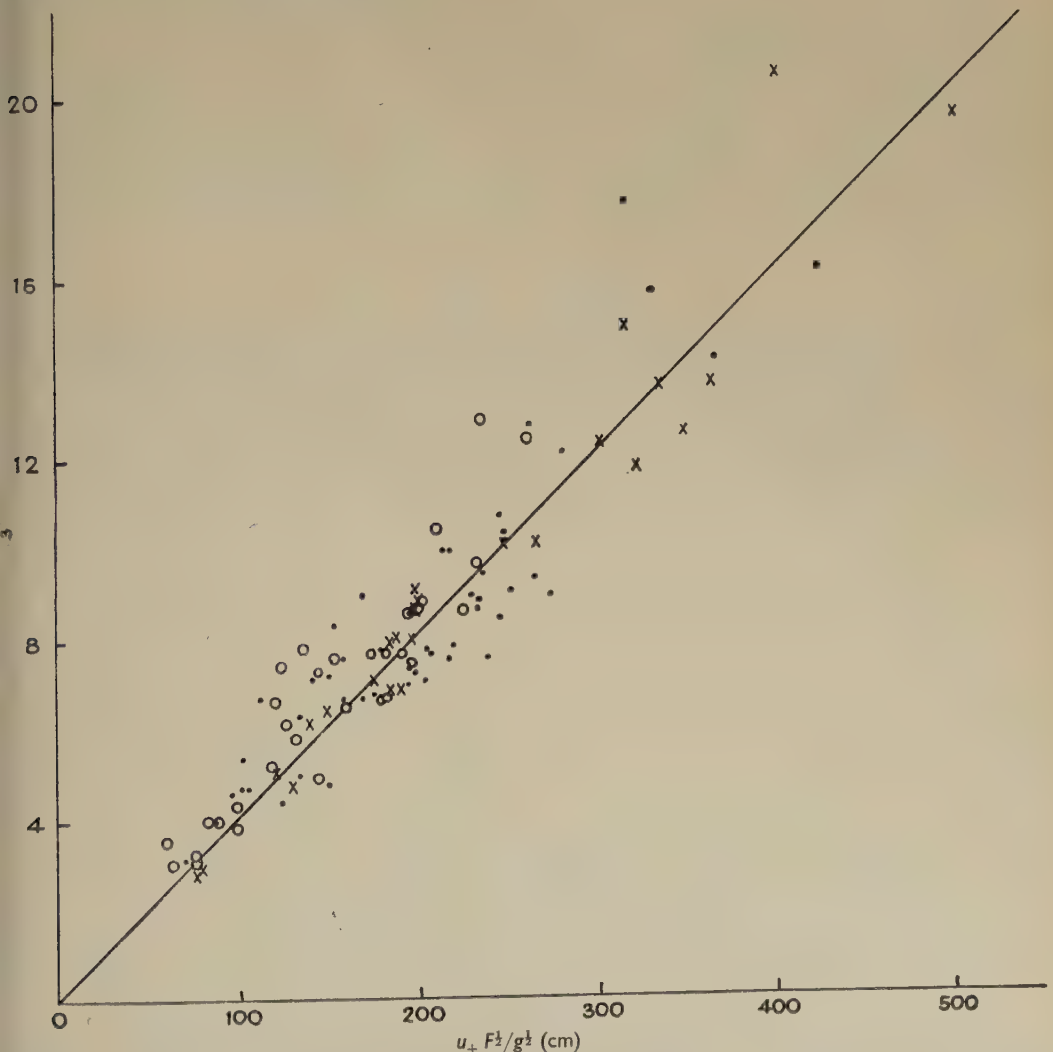

Figure 1. Wave observations by Burling (1954, 1955). The ordinate is $H_{\frac{1}{3}}$, the height exceeded by one in three of the 'dominant' waves ($H_{\frac{1}{3}} \simeq H$).

o 375 < fetch < 600 m; • 600 ⩽ fetch ⩽ 1,000 m; x 1,000 < fetch < 1,350 m.

By contrast the statistics of the wind field are usually confined to a specification of the ' surface wind,' by which is meant the wind at about 10 m above the surface. The tacit assumption is that the atmospheric turbulence, and hence those stresses on the surface which produce the waves, are probably characterized by the mean wind. The windspeed varies with height, almost certainly in such a way that the mean wind at one height is not linearly related to the mean wind at any other. To use the mean wind, even at a standard height, is therefore likely to complicate the resulting formulae; it is the purpose of this note to suggest the use of the friction velocity, a measure of the shearing stress, as a more suitable velocity to characterize the wind field.

It must be well known that it is the stresses on the sea surface produced by the wind, rather than the wind directly, which generate waves. The continued use of the surface wind is probably due to the uncertainty which surrounds the relation between the friction velocity and the wind speed at a given height. Here we shall use a relation for the general form of which there is growing observational support (Charnock 1955, Ellison 1956)

$$\frac{u}{u_+} = \frac{1}{k} \log \frac{gz}{u_+^2} + 11{\cdot}0 \qquad . \qquad . \qquad . \qquad . \qquad (1)$$

Here u is the windspeed at height z, g is the acceleration of gravity, k is von Kármán's constant $\simeq 0{\cdot}4$. The friction velocity, u_+, is defined by $\tau_0 \equiv \rho_a u_+^2$ where τ_0 is the surface shearing stress and ρ_a the density of the air.

u_+ increases more rapidly than $u_{10\,m}$ but precise observations are needed to demonstrate any practical advantage in using it. Suitable observations have been made by Burling (1954, 1955) for short fetch and will be discussed below. At longer fetches the observations are not sufficiently precise but it can be shown that the use of u_+ goes some way toward explaining differences of interpretation which have led to controversy among different workers.

2. Observations at short fetch

If we take it that the equivalent waveheight H is determined by the friction velocity u_+, the fetch F and the acceleration of gravity g then we may expect a relation of the general form $gH/u_+^2 = f(gF/u_+^2)$ (compare Sverdrup and Munk 1947). There is as yet no theoretical method for deriving the form of the function f.

Phillips (1957), as a result of an impressive theoretical paper, has given an expression which determines the waveheight in terms of the pressure fluctuations in the wind and of the time for which the wind has been blowing over an ocean assumed infinite and initially at rest.

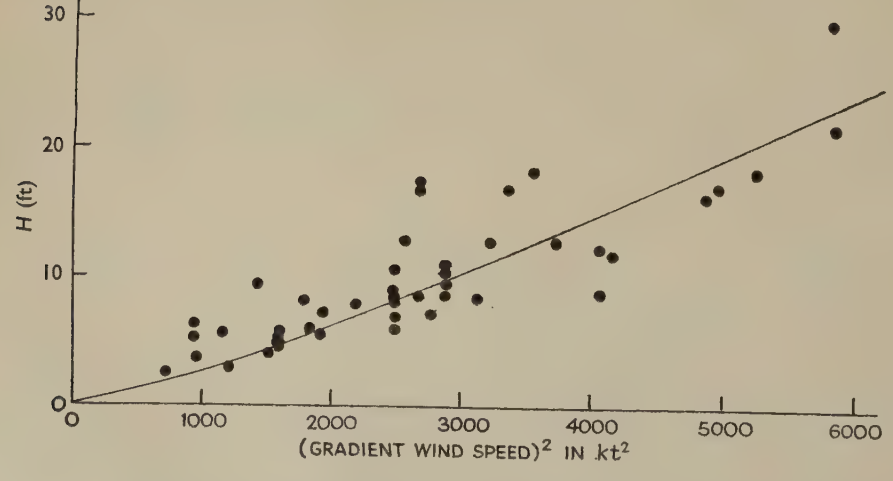

Figure 2. Wave observations reported by Darbyshire (1955). The line indicated is $H = 47\, u_+^2/g$, drawn on the assumption that the gradient windspeed is 3/2 times that at 10 m.

$$\overline{\zeta^2} = \frac{\overline{p'^2}}{2\sqrt{2}\, g\rho_w{}^2 V_c}\, t \qquad . \quad . \quad . \quad . \quad . \quad (2)$$

Here $\overline{\zeta^2}$ is the mean-square fluctuation in water level and $\overline{p'^2}$ the mean-square fluctuation in ne normal pressure on the surface. ρ_w is the density of water, t the elapsed time and V_c is defined y Phillips as the ' convection velocity ' of the surface pressure fluctuations.

To apply Eq. (2) to the finite fetch case it is reasonable, as a first approximation, to make ne following substitutions :

$$\overline{\zeta^2} = 0{\cdot}11\, H^2, \quad \overline{p'^2} = a^2\, \rho_a{}^2\, u_+{}^4, \quad V_c = b^2\, u_+, \quad t = c^2\, F/u_+$$

hen one gets

$$H = 1{\cdot}8\, \frac{ac}{b}\, \frac{\rho_a}{\rho_w}\, \frac{u_+\, F^{\frac{1}{2}}}{g^{\frac{1}{2}}} \qquad . \quad . \quad . \quad . \quad (3)$$

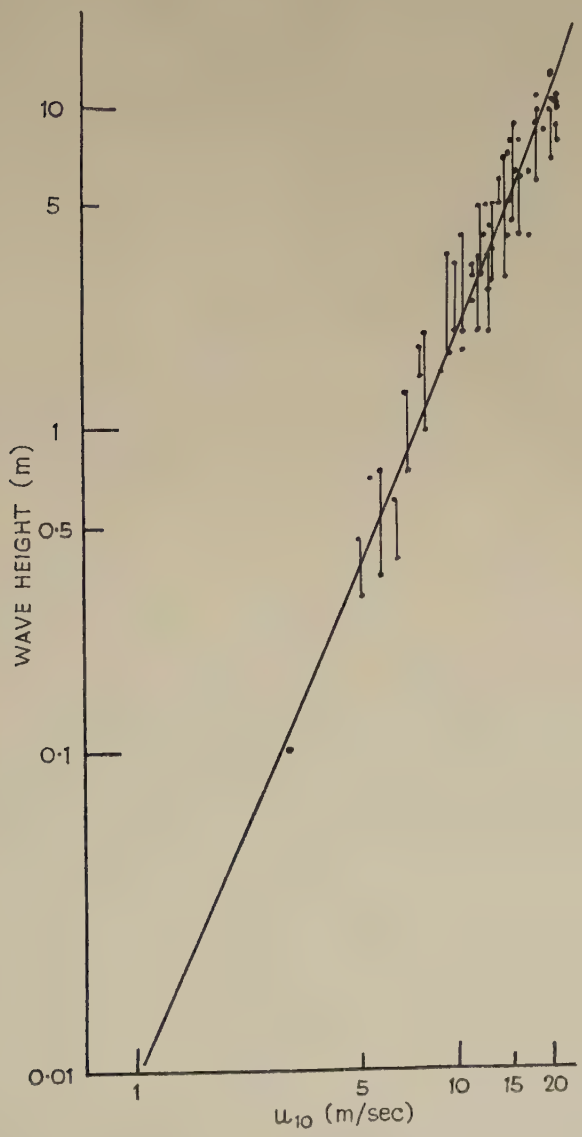

Figure 3. Wave observations reported by Neumann (1953). The ordinate is approximately the ' significant ' waveheight, H_s. $H_s \simeq 1{\cdot}35\, H$. The line drawn is $H = 107\, u_+{}^2/g$.

Burling (1954, 1955) made careful observations of waveheight and wind at short fetch using an adequate wave recorder (Tucker and Charnock 1955) and a calibrated anemometer. His results are shown in Fig. 1 and it will be seen that they are reasonably represented by the expression $H = 0{\cdot}041\ u_+ F^{\frac{1}{2}} g^{-\frac{1}{2}}$. The corresponding value of ac/b is about 19. The constant c is of particular interest and a rough estimate of it can be made by setting $a \simeq 5$ (Charnock unpublished) $b \simeq 5{\cdot}5$ (cf. Eq. (1)), whence $c \simeq 21$. It might be expected that c would be nearer to one and if, following Phillips, we set $a \simeq 95$ then we get $c = 1{\cdot}1$. No observations of pressure fluctuations over a water surface appear to have been reported.

3. Observations at large fetch

At sufficiently large fetch it is reasonable to expect the waveheight to reach an equilibrium value independent of fetch. The technical problems of measuring waves in the open sea have been largely overcome by the development of the shipborne wave recorder (Tucker 1956) but there remain the considerable difficulties of specifying the effective wind and the effective fetch. Darbyshire (1955) presents instrumentally observed waveheights to show that at large fetch the waveheight is proportional to the (windspeed)^2 whereas Neumann (1953) presents largely visual observations to show that the waveheight varies as (windspeed)$^{5/2}$. This difference of opinion has led to some controversy; see, for example, Neumann and Pierson (1957), Darbyshire (1957) and Bretschneider (1957).

If we again take u_+ as being a more relevant variable it turns out that both sets of observations can be represented, within their considerable scatter, by

$$\frac{gH}{u_+^2} = \text{constant} \quad . \quad . \quad . \quad . \quad . \quad (4)$$

This is shown in Figs. 2 and 3 which are used in the forms in which they were originally presented. There remains a considerable discrepancy between the two sets of observations, the constant of Eq. (4) being equal to 47 according to Darbyshire and equal to 107 according to Neumann. At least one set is wrong and confirmatory observations are needed.

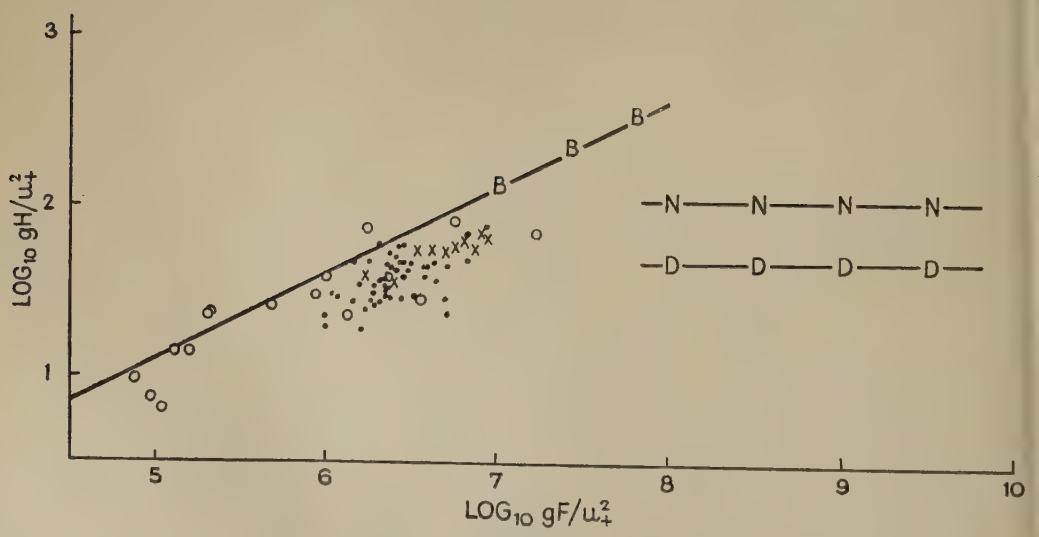

Figure 4. Wave observations reported by Darbyshire (1956). o L. Neagh boat observations, • L. Neagh fixed recorder observations, x Irish Sea observations. The lines indicated are based on the observations of Neumann —— N —— N —— and Darbyshire —— D —— D —— at large fetch and on the observations of Burling —— B —— B —— at short fetch.

4. Observations at intermediate fetch

At intermediate fetch the observations are so imprecise as to be of little use for other than direct engineering applications. Those presented by Darbyshire (1956) are shown in Fig. 4, drawn using logarithmic scales to cope with the large scatter. Those made in winds less than 4 m/sec at 10 m are subject to large errors and have been omitted. The others are not inconsistent with the ideas of Phillips, i.e., $H \propto u_+ F^{\frac{1}{2}}$ up to a limiting value which is probably associated with the mean-square slope of the highest waves. After this the equilibrium result $H \propto u_+^2$ is gradually attained. These results are in fair agreement with those collected by Bretschneider (1952) but there is an obvious need for more precise observations at intermediate and large fetch.

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The emissivity of the earth's surface

By J. T. HOUGHTON

Clarendon Laboratory, Oxford

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Summary

Some measurements from aircraft of the emissivity of the earth's surface are described. They show that for the type of surface studied, the earth's surface is radiating as a black-body at a temperature somewhere between screen air temperature and grass or sea-surface temperature.

1. Measurements

In a previous paper describing the results of measurements from aircraft of the upward and downward fluxes of infra-red radiation in the atmosphere (Brewer and Houghton 1956), it was stated that the measured fluxes of upward radiation were consistently less than would be expected if the earth's surface were a perfect black-body. On these ascents, the chief interest was in the upper atmosphere. The lowest level at which observations of the upward flux was made was 3,000 ft. Also, no measurements were taken of the actual temperature of the ground at places over which the ascent was made, and no record was kept of the exact line of flight on any of these occasions.

More careful observations have been made during February and March 1957, from a Varsity aircraft of the Meteorological Research Flight at Farnborough. Radiometers of the pattern described by Houghton and Brewer (1954) were used. Three areas were chosen for observation : the Farnborough-Basingstoke area consisting mainly of heath land and small forested areas; part of North Cambridgeshire, consisting largely of arable land; and the English Channel near the Royal Sovereign Lightship. Measurements of the upward and downward radiation streams were made after sundown at heights above ground of 200 ft, 1,000 ft and 3,000 ft. Since the time constant of the radiometers is about one minute, the average reading over five minutes of level flight was taken for each level. The observations at each level were repeated, but never showed any significant variation.

All the ascents were made on occasions of clear skies. Under these conditions the downward radiation stream is concentrated in the water-vapour and carbon-dioxide bands. Thus, any reflected by the earth's surface will largely be absorbed by the atmosphere in a short distance and so make little difference to the upward radiation stream at 200 ft above ground. At this level about 60 per cent of the upward radiation stream comes directly from the earth's surface. At 1,000 ft about 50 per cent comes from the earth's surface and at 3,000 ft only about 40 per cent. Thus the measurement at 200 ft should be the most accurate, as far as the emissivity of the earth's surface is concerned.

The results of the ascents are shown in Table 1. The radiation streams F are stated as equivalent temperatures given by $F = \sigma T^4$, where σ is Stefan's constant and T is in °K. The effective temperatures of the earth's surface, shown in column 11, have been calculated from the radiation, water-vapour and temperature data using Yamamoto's radiation chart (Yamamoto 1952). These temperatures have been calculated assuming that the earth's surface is radiating as a black-body at its effective temperature.

On 26 February, patchy fog 10-20 ft thick was lying on the ground. Since liquid water has very strong absorption throughout the infra-red region, it would be expected that fog patches would radiate as black-bodies, and obscure the emission from the ground itself. During the ascents described, an observation was made just above a layer of stratus cloud at about 3,000 ft. The same result was obtained as found previously by Brewer and Houghton (1956), that the emissivity of such cloud is 100 per cent.

TABLE 1. COMPARISON BETWEEN GROUND AND AIRBORNE MEASUREMENTS

		Ground measurements				Airborne measurements						
1	2	3	4	5	6	7	8	9	10	11	12	13
Date and time (GMT)	Station	Mean screen air temp. (°C) 1800-1900	Change in air temp. (°C) 1800-1900	Mean ground or sea temp. (°C) 1800-1900	Change in ground temp. (°C) 1800-1900	Height above ground (ft)	Air temp. (°C)	Dew point (°C)	Equiv. temp. (°C) upward radiation stream	Effective temp. (°C) of earth's surface (calculated)	Screen air temp. (°C) space-time average (mean of column 3)	Ground or sea temp. (°C) space-time average (mean of column 5)
26 Feb. 1957 1815-1845	South Farnborough	6·7	− 1·1	1·7	− 2·8	200	7·6	3·0	5·5	4·4	6·4	1·1
	Odiham	6·1	− 1·1			1,000	6·5	0·0	5·0	3·0		
	Basingstoke	6·5	− 1·1	0·5	− 2·2	3,000	1·6	− 4·5	2·8	3·2		
28 Feb. 1957 1815-1845	South Farnborough	5·2	− 0·2	2·6	− 0·4	200	4·5	− 1·5	3·8	3·2	4·7	2·7
	Odiham	4·3	− 0·3			1,000	3·0	− 3·0	3·6	3·8		
	Basingstoke	4·7	− 0·1	2·8	0·0	1,250	2·4					
						1,750	8·4					
						3,000	7·4	− 39	5·3	3·4		
28 Feb. 1957 1915-1945	Royal Sovereign Lightship	6·1 (1900-2000)	− 0·1 (1900-2000)	6·7 (1900-2000)	0·0 (1900-2000)	200	6·9	0	7·6	8·1	6·1	6·7
						1,000	5·3	− 3	6·8	7·8		
						2,500	8·3	− 32	7·7	8·0		
13 Mar. 1957 1825-1855	Feltwell	12·2	− 0·2	8·9	0·0	200	12·6	7·8	12·5	12·4	12·4	9·6
	Mildenhall	12·8	− 0·7	11·1	− 1·1	1,000	11·7	7·5	12·4	12·9		
	Waterbeach	12·2	− 1·0	8·8	− 0·4	3,000	9·9	0·0	10·6	11·2		

On 28 February, 6/8 cirrus cloud was present. Since stars could be seen through this cloud it was thin, and again referring to the previous measurements of Brewer and Houghton (1956) it should make little difference to the radiation field at the ground. This was confirmed by measurement from the aircraft of the downward radiation stream.

The uncertainty in the readings of the downward-facing radiometer is about $\frac{1}{2}$°C for the first two ascents, and about 1°C for the ascent on 13 March. This possible error will be reflected in the effective temperature of the ground.

2. Discussion of results

On comparing the effective temperature of the ground calculated from the airborne measurements as shown in column 11, with the observations of ground temperature and screen-air temperature shown in columns 12 and 13, it will be seen that, for the observations over land, the radiation emitted by the earth's surface corresponds to an effective temperature somewhere between the readings of an air-temperature thermometer and a grass thermometer. This is the result one would expect since some of the radiation will come from foliage and other material mainly at air temperature, and some will come from the ground itself. We thus deduce that for the type of surface studied, the earth is radiating effectively as a black body. The one set of observations which has been made over the sea gives a value for the effective temperature about 2°C greater than either the screen air temperature or the one observation of the sea-surface temperature made at the time. Since so few observations were made no significance can be given to this difference. Again, the measurements are consistent with an emissivity for the sea surface of 100 per cent with a possible error of less than 5 per cent. An error of 1°C in effective temperature implies an error in the emissivity deduced of just over 1 per cent.

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The washout of lycopodium spores by rain

By F. G. MAY

United Kingdom Atomic Energy Authority Research Group, Harwell

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SUMMARY

This report describes the determination of the washout coefficient of lycopodium spores by rain. The experimental results are in good agreement with those obtained theoretically by Chamberlain (1953) after allowance has been made for the varying raindrop spectra of different types of rain.

1. INTRODUCTION

The washout by rain of aerosols in the atmosphere is of considerable interest in the field of atomic energy. A theoretical treatment of the subject was given by Chamberlain (1953), who calculated the washout coefficients of a number of aerosols at several rates of rainfall. The experiments described below were carried out to check the validity of the theory for a specific aerosol.

Lycopodium spores were chosen as the aerosol because they could be conveniently liberated as a monodisperse aerosol marked with I^{131}, and they had a terminal velocity, 2 cm/sec, within the range considered in the theoretical treatment.

2. METHOD

Lycopodium spores, marked with I^{131}, were released from a height of 3·3 m during a period of rain. Samples of the rain were taken along a 10 m radius arc, downwind, at ground level. Direct deposition was measured using similar samplers from which the rain was excluded. The difference in activity of the two samples was due, therefore, to collection of the spores by the raindrops in falling through the aerosol cloud. The washout coefficient, i.e., the proportion of the aerosol cloud washed out per second, was then calculated. The height of release was limited to 3·3 m for ease of construction and accessibility. The ratio 1 : 3 of height to sampling distance was chosen as the best compromise between the errors due to direct deposition by downwind turbulent diffusion, and to non-collection of the smallest raindrops by their horizontal displacement in the wind.

3. EXPERIMENTAL DETAILS

(*a*) *Spores*

The spores were marked by separating 100 mc of carrier-free NaI^{131}, using nitrous acid, into 50 ml of carbon tetrachloride, and pouring this solution over 5-10 g of spores, which were then dried and placed in the dispersing bottle. The activity on the spores was measured by gamma counting at known distances from a GM-4 Geiger tube in a special rig which had originally been calibrated using a radium source of known strength.

(*b*) *Dispersing tower*

The disperser was then mounted on a windvane surmounting a 3 m high tower and connexions made, by rubber tubing, to a compressed-air supply and a mercury manometer located on the tower. Contacts were arranged in the manometer so that warning was given if the pressure rose above, or fell below, the working pressure (20 cm Hg). The disperser was vibrated during release of the spores by means of a modified electric bell, which was also mounted on the windvane.

Beneath the windvane was located an ion chamber, which gave a continuous measure of the activity remaining in the disperser. The rate of release of the spores could thus be checked and if any change occurred in the rate of rainfall during the experiment a weighted mean value for the rainfall rate could be calculated.

(c) *Sampling*

The original sampling arrangement was a ring of crystallizing basins, spaced 1 metre apart on a circle radius 10 m, centred at the 3 m high tower from which the spores were liberated. The basins were 10 cm in diameter and were set in level sockets embedded in concrete. The shape of the collecting vessels was chosen so that loss by splashing would be minimized.

Preliminary trials in dry weather indicated no appreciable fallout of spores in the collectors and it was thought that the vertical diffusion was insufficient to carry the spores from a height of 3·3 m to ground level in a travel of 10 m. It was therefore assumed that the spores collected in trials done during rain had been collected by raindrops. Later this assumption was reconsidered and further experiments on dispersal from the tower were then carried out, in both dry and wet weather, using smoke candles.

It was found that the turbulence varied considerably from day to day. On fine sunny days the smoke cloud usually did not reach ground level for 30-50 m, but on dull rainy days it did so much more quickly, often inside the ring of samplers. It was therefore possible for the aerosol cloud to be deposited in the samplers by turbulent diffusion. The sampling arrangement was then altered to measure the direct deposition to the samplers as well as the total deposition. Samplers shielded from rain were installed midway between each pair of rain collectors. Each of these consisted of a retort stand, set in concrete, on the base of which was a 10 cm diameter crystallizing basin. Ten cm above the rim of the basin was a 30 × 40 cm horizontal aluminium plate, which allowed free passage of air and airborne spores over the basin but prevented the entry of rain. It was not possible to determine whether this arrangement affected the direct deposition in the control basins by dry-weather experiments from the tower, as the scale of turbulence appeared to be quite different in dry and wet weather. Previous experience indicated that there would be little difference in deposition in the covered and uncovered basins from a ground-level release in dry weather.

This method was applied to the last five field experiments, and enabled the deposition by turbulent transfer to be estimated and subtracted from the total deposition to give the deposition due to washout.

(d) *Meteorological measurements*

The windspeed was measured by means of a recording cup anemometer upon a second tower 25 m away at the same height, 3·3 m, as the disperser.

The rate of rainfall was measured by a tilting-siphon rain-gauge, also 25 m away. The clock mechanism gave one revolution of the chart every hundred minutes. It was possible to read the rate of rainfall over periods of a few minutes.

(e) *Method of operation*

During operation of the disperser the reading of the ion-chamber was noted every 30 sec until either it remained constant or the rain ceased. The rain water and direct deposition samples were collected, the basins being washed out with a 10 per cent caustic soda solution to which a little KI had been added. The activity remaining in the disperser was estimated as before, using gamma-counting.

The duration of each release varied between 10 and 30 min.

(f) *Counting*

The caustic solution was first filtered to remove dust, and the residue dried and beta-counted. 75 ml of 10 per cent NaI solution were added to the filtrate, which was then neutralized with HNO_3, and the iodine precipitated as silver iodide by the addition of silver nitrate. This was filtered in a special light-tight filter stick to reduce decomposition. The silver iodide was dried, weighed, and beta-counted as an infinitely thick source to give the specific activity of the silver iodide. The source holder had an adjustable bottom so that, regardless of the quantity of silver iodide obtained, the upper surface of the source could be brought level with the brim of the tray. Thus the counting geometry was exactly reproducible. The minimum weight of silver iodide required to give an infinitely thick source was 10 g. The total activity in the sample was found by multiplying the weight of silver iodide by the specific activity and adding a small correction for the activity on the residue.

4. Treatment of results

If the total activity found in the rain-water samples is C micro micro curies, the activity in the covered samples is C_0 micro micro curies, and the exposed area of each sampler A cm^2, then the total integrated cross-wind deposition (the deposition per cm travel of the aerosol cloud)

$$I_t = \frac{100\,C}{A}\ \mu\mu\text{c/cm} \qquad (1)$$

since the samplers are 100 cm apart.

Similarly the integrated cross-wind deposition due to direct deposition

$$I_d = \frac{100\,C_0}{A}\ \mu\mu\text{c/cm} \qquad (2)$$

and the integrated cross-wind deposition due to washout

$$I_w = I_t - I_d = \frac{(C - C_0)\,100}{A}\ \mu\mu\text{c/cm} \qquad (3)$$

The proportion of the aerosol cloud deposited per second is

$$\Lambda_t = \frac{I_t\,U}{10^9\,Y} = \frac{10^{-7}\,CU}{AY}\ \text{sec}^{-1} \qquad (4)$$

where Y is the total activity released, in millicuries, and U is the wind velocity in cm/second.

Similarly the proportion of the aerosol cloud deposited by washout per second is

$$\Lambda_w = \frac{I_w\,U}{10^9\,Y} = \frac{10^{-7}\,(C - C_0)\,U}{AY}\ \text{sec}^{-1} \qquad (5)$$

5. Results

The results are given in Table 1. Λ_t and Λ_w are compared with the theoretical washout coefficient Λ_{theo} computed by Chamberlain (1953) for an aerosol of 2 cm/sec terminal velocity (Fig. 1).

Chamberlain derived the washout coefficient in the following manner.

If $M_{(s)}\,ds$ is the concentration of water in the air, in g/cm^3, composed of drops whose radii lie between s and $s + ds$, and $N_{(s)}\,ds$ is the number of such raindrops :

$$N_{(s)}\,ds = \frac{3}{4\pi s^3}\,M_{(s)}\,ds \qquad (6)$$

Then, if $E_{(r,\,s)}$ is the impaction efficiency of a raindrop radius s falling with its terminal velocity $V_{(s)}$ for an aerosol of radius r, the number of drops with radius between s and $s + ds$ falling per second through each cm^2 perpendicular to the direction of rainfall is

$$N_{(s)}\,V_{(s)}\,ds = \frac{3}{4\pi s^3}\,M_{(s)}\,V_{(s)}\,ds \qquad (7)$$

and the proportion of a monodisperse aerosol, radius r, removed per second by these drops is

$$\lambda_{(r,\,s)}\,ds = \pi s^2\,N_{(s)}\,V_{(s)}\,E_{(r,\,s)}\,ds \qquad (8)$$

Hence the proportion of the aerosol deposited per second by rain is

$$\Lambda_{\text{theo}} = \int_{s=0}^{s=s_{\max}} \lambda_{(r,\,s)}\,ds$$

$$= \int_{s=0}^{s=s_{\max}} \frac{3}{4s}\,M_{(s)}\,V_{(s)}\,E_{(r,\,s)}\,ds \qquad (9)$$

Chamberlain evaluated Λ_{theo} for various rates of rainfall, using Best's figures (1950) for $M_{(s)}$ and $V_{(s)}$, Langmuir's figures (1948) and Langmuir and Blodgett's figures (1949) for $E_{(r,\,s)}$, (Table 2).

TABLE 1. SUMMARY OF RESULTS

Date	Windspeed (cm/sec)	Rate of rainfall (mm/hr)	Total activity released (mc)	I_t Total (10^3 $\mu\mu$c/cm)	I_d Direct deposition (10^3 $\mu\mu$c/cm)	I_w Washout (10^3 $\mu\mu$c/cm)	Proportion of active cloud directly deposited in samplers (10^{-7})	Λ_t (10^{-4}/sec)	Λ_w (10^{-4}/sec)	Λ_{theo} Chamberlain using $E_{(r,s)}$ Langmuir (10^{-4}/sec)	Type of rain
16.7.53	500	0·10	9·6	2·33	not measured	not measured	1·05*	1·21	not measured	0·5	drizzle
22.7.53	570	1·65	18·0	33·9	,,	,,	7·1*	10·7	,,	5·2	frontal rain
5.5.54	361	23	4·30	95·1	,,	,,	83·9*	80	,,	39	thunderstorm
20.8.54	202	23	5·90	92·6	,,	,,	—	36·7	,,	39	thunderstorm
5.4.56	490	0·66	23·7	12·3	,,	,,	0·42*	2·56	,,	2·28	shower
12.4.56	341	1·29	14·2	34·4	,,	,,	9·1*	8·3	,,	4·1	frontal rain
9.5.56	854	2·4	29·3	45·7	,,	,,	5·7*	13·3	,,	6·7	frontal rain
24.5.56	412	0·06	10·2	7·4	,,	,,	4·8*	3·0	,,	0·3	drizzle
29.5.56	264	0·74	17·4	28·1	,,	,,	5·1*	4·26	,,	2·45	shower
7.6.56	548	0·40	11·6	2·64	,,	,,	—	1·25	,,	1·35	shower
12.6.56	434	0·20	16·7	1·73	,,	,,	—	0·45	,,	0·64	shower
16.8.56	320	3·91	7·88	27·6	2·4	25·2	2·25	11·2	10·2	9·72	frontal rain
27.9.56	543	1·12	19·9	17·9	2·56	15·3	0·95	4·88	4·17	3·63	frontal rain
25.10.56	845	14·1	40·8	183	33·9	149	6·1	37·9	30·8	26·8	heavy frontal rain
11.12.56	334	1·01	32·8	32·7	1·38	31·3	0·31	3·3	3·2	3·23	frontal rain
31.12.56	332	3·64	4·58	14·3	1·99	12·3	3·2	10·4	8·9	9·2	continuous rain of showery type

* These values not obtained experimentally but deduced as the correction necessary to convert Λ_t to Λ_{theo}.

TABLE 2. COMPUTATION OF A_{theo}

$$A_{theo} = \int_{s=0}^{s=s_{max}} \frac{3}{4s} M_{(s)} V_{(s)} E_{(r,s)} \, ds$$

Radius of raindrops s 10^{-2} cm)	Terminal velocity $V_{(s)}$ (cm/sec)	Impaction efficiency $E_{(r,s)}$ Langmuir	Impaction efficiency $E_{(r,s)}$ Mason	Electrostatic collection efficiency for uncharged raindrops	$M_{(s)}$ in 10^{-9} g/cm^3 Concentration of water composed of drops between $s - \delta s$ and $s + \delta s$ where $\delta s = 6{\cdot}25 \times 10^{-3}$ cm Rate of rainfall 0·5 mm/hr	1·0 mm/hr	2·5 mm/hr	5 mm/hr	10 mm/hr	25 mm/hr
0·625	37	0·66	1·12	2·50	1·3	1·6	2·2	2·7	3·4	4·6
1·875	148	0·80	0·98	0·66	4·4	5·7	7·9	10·1	12·7	17·3
3·125	248	0·85	0·96	0·38	6·9	9·5	13·8	17·9	23·1	31·8
4·375	337	0·89	0·96	0·26	7·7	11·6	18·3	24·8	32·9	46·4
5·625	425	0·90	0·95	0·18	6·8	11·7	20·6	29·7	41·0	59·9
6·875	500	0·91	0·95	0·13	4·8	10·0	20·6	31·9	46·5	71·3
8·125	565	0·91	0·94	0·10	2·9	7·3	18·4	31·5	49·0	79·8
9·375	618	0·91	0·94	0·08	1·4	4·7	14·9	28·7	48·6	84·9
10·625	667	0·91	0·93	0·06	0·6	2·6	11·1	24·5	45·5	86·5
11·875	707	0·92	0·93	0·05	0·2	1·3	7·5	19·4	41·4	84·6
13·125	742	0·92	0·93	0·05	0·1	0·6	4·6	14·4	34·1	79·9
14·375	774	0·92	0·92	0·04		0·2	2·6	10·1	27·5	73·0
15·625	803	0·91	0·92	0·04		0·1	1·4	6·6	21·1	64·3
16·875	827	0·91	0·91	0·03			0·7	4·0	15·4	55·0
18·125	846	0·90	0·90	0·03			0·3	2·4	10·8	45·6
19·375	863	0·89	0·90	0·03			0·1	1·3	7·2	36·6
20·625	876	0·89	0·89	0·02				0·6	4·6	28·5
21·875	887	0·88	0·89	0·02				0·3	2·8	21·6
23·125	896	0·88	0·89	0·02				0·1	1·6	15·8
24·375	903	0·87	0·88	0·02				0·1	0·9	11·2
25·625	911	0·86	0·88	0·02					0·5	7·8
26·875	916	0·86	0·87	0·02					0·2	5·2
28·125	921	0·85	0·87	0·01					0·1	3·4
29·375	925	0·85	0·86	0·01					0·1	2·1
30·625	929	0·84	0·86	0·01						1·3
31·875	931	0·84	0·85	0·01						0·8
33·125	933	0·83	0·85	0·01						0·4
34·375	935	0·83	0·84	0·01						0·3
35·625	936	0·82	0·84	0·01						0·1
			Mason A_{theo}		$2{\cdot}00 \times 10^{-4}$	$3{\cdot}52 \times 10^{-4}$	$7{\cdot}30 \times 10^{-4}$	$12{\cdot}6 \times 10^{-4}$	$21{\cdot}5 \times 10^{-4}$	$42{\cdot}8 \times 10^{-4}$
			Langmuir A_{theo}		$1{\cdot}82 \times 10^{-4}$	$3{\cdot}24 \times 10^{-4}$	$6{\cdot}84 \times 10^{-4}$	$11{\cdot}9 \times 10^{-4}$	$20{\cdot}6 \times 10^{-4}$	$41{\cdot}3 \times 10^{-4}$
Calculated proportion of spore cloud captured by uncharged drops per second by induced electrostatic collection					$0{\cdot}71 \times 10^{-4}$	$1{\cdot}05 \times 10^{-4}$	$1{\cdot}72 \times 10^{-4}$	$2{\cdot}46 \times 10^{-4}$	$3{\cdot}50 \times 10^{-4}$	$5{\cdot}50 \times 10^{-4}$

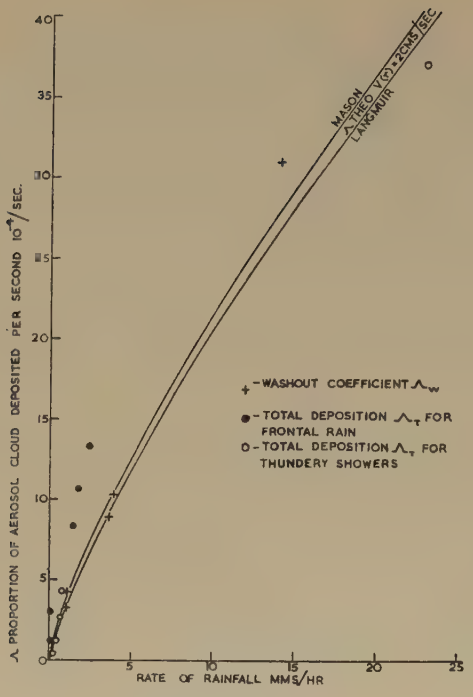
Λ PROPORTION OF AEROSOL CLOUD DEPOSITED PER SECOND 10⁻⁴/SEC.
RATE OF RAINFALL MMS/HR
MASON
Λ THEO V(s) = 2CMS/SEC
LANGMUIR
+ - WASHOUT COEFFICIENT Λw
● - TOTAL DEPOSITION Λt FOR FRONTAL RAIN
O - TOTAL DEPOSITION Λt FOR THUNDERY SHOWERS

Figure 1. Variation of Λ with rate of rainfall.

6. Discussion of results

(a) *Direct deposition*

Eleven experiments were carried out before it was discovered that the aerosol cloud was occasionally reaching ground level within the sampling ring. If it is assumed that the difference between Λ_t of these experiments and Λ_{theo} is entirely due to direct deposition, the proportion of the aerosol cloud directly deposited in the samplers can be calculated. Upon comparing these figures with the known proportion directly deposited in the final set of experiments fair agreement is found in all but one result (Table 1).

(b) *Variation in drop-size spectrum*

Best's figures of $M_{(s)}$ for various rates of rainfall are based on the average of a large number of observations from all over the world. However, the raindrop spectrum depends not only on the rate of rainfall but also on the type of rain and, in thundery showers, the position of the storm centre relative to the sampling point. The spectrum for thundery showers tends to be narrower, and is usually composed of much larger drops, than the spectrum for frontal rain of the same intensity (Best 1950).

If for the sake of simplicity it is assumed that the rain is composed of drops of all the same size, then at a given rate of rainfall, $M_{(s)}$ is inversely proportional to $V_{(s)}$ since $M_{(s)} \cdot V_{(s)}$ is equal to the mass of rain falling on one cm^2 per second. Hence from Eq. 9 the washout coefficient is almost inversely proportional to the radius of the raindrops s, since the impaction efficiency $E_{(r,s)}$ varies little with the different values of s.

Since rain is composed of drops of different sizes this simple picture of washout does not hold true, but it does indicate that the washout coefficient for frontal rain should be slightly higher than Λ_{theo} and the coefficient for thundery showers should be less than Λ_{theo}. This is borne out by the results for frontal rain at least.

(c) *Interception*

Langmuir in calculating the impaction efficiency $E_{(r,s)}$ for a raindrop falling through an aerosol cloud assumed that each aerosol particle could be considered as a point of negligible width. However, the edge of the aerosol particle may come into contact with the edge of the raindrop and in such a case, if the aerosol is easily wetted, it will be collected by the raindrop. In this case for an aerosol of comparable size with the raindrop Langmuir's figures will underestimate the collection efficiency.

Mason has recently calculated the impaction efficiency of raindrops allowing for interception. Theoretical washout coefficients calculated from both Mason's and Langmuir's figures are given in Table 2.

(d) *Electrostatic collection*

It was found experimentally that the spores carried charges of either sign up to a maximum of about 6×10^{-4} e.s.u. The average charge per spore of either sign was estimated to be $2{\cdot}5 \times 10^{-5}$ e.s.u. and the average of the squares of the charges to be 35×10^{-10} (e.s.u.)^2 per spore.

No measurements of the charge on raindrops were carried out in the experiments, but from work by Smith (1955) it appears that the average and the maximum charges per drop are dependent on the drop size and the local electric field. The average charges per drop vary enormously, from $1{\cdot}2 \times 10^{-1}$ e.s.u. for 3 mm diameter drops in a thunderstorm, down to less than 5×10^{-5} e.s.u. for certain periods of frontal rain. Not all frontal rain is lightly charged, but it is true to say that frontal rain is not usually so heavily charged as thundery rain.

Electrostatic collection of charged spores is due to two effects; the coulombic attraction between oppositely charged spore and raindrop, and the induced attraction due to the polarisation of a raindrop in proximity to a charged spore. Using the results of Kraemer and Johnstone (1955) it can be calculated that in very highly charged rain (average charge per drop of the order of $2{\cdot}5 \times 10^{-1}$ e.s.u.) the total coulombic capture would be equal to the collection due to impaction, and the washout coefficient would be doubled. From Smith's data it is apparent that these conditions are extremely rare although not impossible. Smith occasionally observed average charges of about $3{\cdot}0 \times 10^{-2}$ e.s.u. per drop in thunderstorms, and under these conditions the coulombic collection is only of the same order as the induced electrostatic capture, i.e. 10 per cent of the total washout. Since the average charge per drop for all rain was found by Chalmers and Pasquill (1938) to be about 10^{-3} e.s.u. per drop it appears that the coulombic collection can usually be neglected, for frontal rain at least.

The induced electrostatic capture has been calculated in Table 2 for spores of mean-square-charge 35×10^{-10} (e.s.u.)^2 per spore in various rates of rainfall using the data of Kraemer and Johnstone (1955).

It is interesting to note that the largest contribution to the induced electrostatic capture is made by the small droplets. This is also true for the coulombic capture in spite of the fact that the large drops carry much greater charges than the small drops. This is due, in both cases, partly to the much greater numbers of the small drops and partly to the fact that, since the small drops have a very low terminal velocity, the spores remain in their electric field for a much longer period.

(e) *Non-collection of raindrops*

The terminal velocity of the smallest raindrops may be insufficient to bring them to ground level within the sampling ring after passing through the aerosol cloud. The error so caused is small except when the rate of rainfall is low and the windspeed high. At normal windspeeds the contribution to the washout which is lost by non-collection of the smallest drops tends to balance the increase due to electrostatic capture. Moreover the loss of the smallest drops reduces the electrostatic capture considerably (see Table 2).

In Table 3 the combined corrections for induced electrostatic capture in lightly charged rain and for non-collection of raindrops are given as percentages of the theoretical impaction washout coefficients. This was done by taking a windspeed such that all raindrops below the first, second, third or fourth size groups will fall outside the sampling ring after passing through the aerosol cloud. The proportion of the aerosol cloud electrostatically captured per second by drops larger than the size selected is estimated and from this figure is subtracted the proportion of the aerosol cloud captured per second on impaction and interception by the drops smaller than the selected size. The resultant figure is then given as a percentage of the theoretical washout coefficient using Mason's figures for the impaction efficiency.

TABLE 3. ESTIMATED POSSIBLE ERROR IN EXPERIMENTAL RESULTS DUE TO NON-COLLECTION OF SMALLER RAINDROPS AND INDUCED ELECTROSTATIC CAPTURE IN LIGHTLY CHARGED RAIN

Windspeed (cm/sec)	Rates of rainfall					
	0·5 mm/hr	1·0 mm/hr	2·5 mm/hr	5·0 mm/hr	10 mm/hr	25 mm/hr
	Estimated possible error (per cent)					
0	+ 36	+ 30	+ 24	+ 20	+ 16	+ 13
270	+ 26	+ 22	+ 19	+ 16	+ 14	+ 11
600	+ 4	+ 7	+ 8	+ 8	+ 8	+ 7
876	− 24	− 15	− 7	− 3	− 1	+ 1

7. EXPERIMENTAL ERROR

The spores were gamma counted before and after release to an accuracy of rather better than 1 per cent. The activity released was known therefore to an accuracy of 1 per cent. The activity of the rainwater samples could be measured to an accuracy of 3 per cent. The accuracy of the wind velocity measurements was probably of the order of 4 per cent. Thus the probable accuracy of each of the results for Λ_w is about 8 per cent. The rate of rainfall could be measured with an accuracy between 5 and 10 per cent. These estimates of the experimental accuracy do not include systematic errors such as the non-collection of raindrops, which have been referred to elsewhere.

8. CONCLUSIONS

The results of this series of experiments indicate that Chamberlain's figures of the washout coefficient for particles with a terminal velocity of 2 cm/sec are substantially correct. In frontal rain, however, the washout coefficient will probably be slightly higher than that predicted by Chamberlain, and in showers it is likely to be lower.

It is considered unlikely that any aerosol, in this size range, produced as a result of a reactor accident would be so heavily charged as the lycopodium spores used in these experiments. Thus deposition by electrostatic collection is unlikely to play an important part in the dilution of the aerosol cloud, even for freshly released aerosols.

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Temperature variations at Resolute, Northwest Territories

By RICHMOND W. LONGLEY
Meteorological Branch, Department of Transport, Toronto

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Summary

From a study of temperature data at Resolute, Northwest Territories, it is shown that in the Arctic regions the time of minimum temperature occurs near the end of February or in early March. This delay, relative to the corresponding time in temperate latitudes, is a result of the lack of incoming radiation during the polar night. Because of fog off open water in late July, the maximum temperature occurs during the second week of July.

1. Introduction

The ship *Alert* spent the winter of 1875-1876 off the northeast coast of Ellesmere Island. In describing his experiences during the expedition, Markham (1884) remarked, ' Strange to say, our extreme cold came with the returning sun. During the latter end of February the temperature had been gradually getting lower and lower, until on the 4th March it had fallen to — 74° . . . For the week ending March 4th the mean temperature was — 60°.'

Markham's surprise has been matched by other people who have wintered in the polar regions. These, accustomed to moderating temperatures by the end of February, expect that the return of the sun early in the month will herald the end not only of the period of darkness but also of the worst of the low temperatures.

The annual trends of temperature have been the subject of many studies. These have shown that the times of minimum temperature vary from mid-January for continental locations to early February for maritime districts. They also have shown that the results of fitting a sine curve to the monthly mean temperatures give an accurate description of the trend. When people have considered the subject in relation to polar regions they have assumed that here too the annual trend could be obtained through the use of the monthly means. Approximately a decade of observations at Resolute, Northwest Territories (75°N, 95°W), now makes a study possible although conclusions must still be tentative since non-cyclical variations still obscure the trends.

2. Annual trends of temperature

The annual trend of temperature has its source in the change of heat received from the sun. For Resolute and other places within the Arctic circle, the use of monthly means leads to errors in the derived annual trend. In these regions the variation in the incoming radiation is smooth only from the time of the sun's first appearance in mid-winter until it last sets in the autumn. During the remaining weeks of the year the incoming radiation is zero. During March the temperature rises rapidly, giving a higher mean for the month than for January. A sine curve through the means gives a minimum early in February. Yet at Resolute the sun first appears on 5 February, and the incoming radiation is negligible for two or three weeks after this date.

The situation is similar to the diurnal trend in more southerly latitudes. The incoming radiation is cut off when the sun sets and the minimum temperature occurs about one half-hour after sunrise, at which time incoming and outgoing radiation are equal. Similarly at stations inside the Arctic circle the minimum temperature for the year must occur after the sun first rises in the late winter, at that time when incoming equals outgoing radiation.

Information on the annual trend of temperatures for Resolute was obtained by studying the available temperature data for the period from October 1947, when the station was established, to March 1957. Data for 16 months of the period were lacking and so the results are based on eight or nine years of observations. The mean maximum and minimum temperatures for each day were computed from these, and then the seven-day running mean temperatures were calculated. The computation of the running mean smooths somewhat the irregularities in the trend. The resultant curves are shown in Fig. 1. Also plotted are the mean monthly maximum and minimum temperatures.

The curves in Fig. 1 show that the non-cyclical variations have not been completely eliminated. Thus there is a dip in the curves near the end of March, and another during the second half of

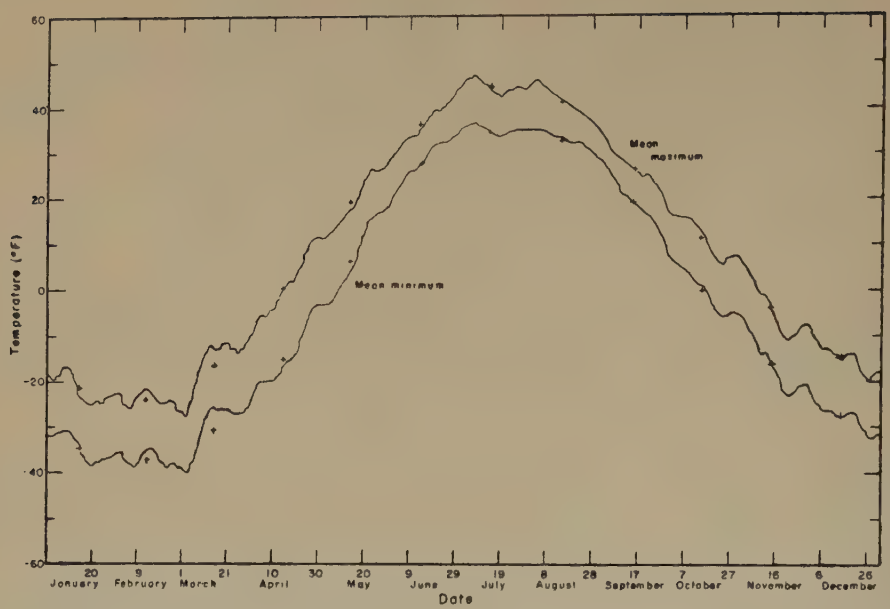

Figure 1. Mean temperature, Resolute, N.W.T. Eight years' data. Lines give seven-day running mean, and +'s monthly, mean maximum and minimum temperatures.

July. By mentally smoothing the curves during the period 15 March to 15 October it is discovered that the maximum deviation from the smooth trend is approximately 5° (all temperatures are quoted in degrees Fahrenheit).

The week of the year which has the maximum mean temperature is that centred on 8 July, with a mean maximum of 46·9° and a mean minimum of 36·4°. In view of the maritime location of Resolute, this time is early, indicating a lag of only 18 days after the summer solstice. The general curve tends to suggest that the non-cyclical effects have caused this time to be early. The mean for 3 August, a secondary maximum, is only 1·2° below that for 8 July. Between these two dates there is a secondary minimum on 19 July, 3·7° below the 8 July value. If this minimum is a result of non-cyclical cool spells, a correction of 5° would bring the value for the second half of July above the calculated mean temperature of 41·6° for 8 July. We may conclude that the warmest time of the year lies somewhere between 7 July and 4 August, with a probable time between 8 July and 20 July.

It seemed desirable to learn whether these irregularities are the results of chance variations. The mean maximum was determined for each day by drawing a smooth curve to follow the trend in the seven-day running mean temperatures. This curve was drawn to show a minimum on 1 March and a maximum on 22 July. Deviations of the observed means from these were computed. (In these computations data to April 1958 were used). Then the standard deviations of these departures were computed by months. Similarly a smoothed curve for the minima, showing a minimum on 4 March and a maximum on 20 July, was used to compute the monthly standard deviations of departures of the minimum temperatures. These values are given in Table 1.

TABLE 1. Standard deviations of the departures of temperature from smoothed curves (in degrees Fahrenheit)

	S.D. for max. (°F)	S.D. for min. (°F)		S.D. for max. (°F)	S.D. for min. (°F)
Jan.	11·5	9·8	July	5·8	3·9
Feb.	11·6	9·9	Aug.	5·6	3·4
Mar.	12·0	11·5	Sept.	4·8	5·9
Apr.	10·5	11·7	Oct.	8·8	9·9
May	7·3	8·4	Nov.	11·8	11·9
June	4·3	4·1	Dec.	11·0	9·6

According to Fig. 1, there is a peak in the seven-day running mean temperature about 5 March. For 13 March, the average departure of 10 minimum temperatures from the smoothed mean of − 32·0° is 11·6°. This is almost exactly three times the standard error of the mean of 10 observations, so the value has theoretically one chance in 100 of arising by pure chance. Similarly it was calculated that for 12 July the average departure of the minimum for 9 days was 3·6°, which is 2·6 times its standard error. These were the times when the departures relative to their standard errors were the greatest. Therefore the deviations from smoothed curves following the curves of Fig. 1 can be assumed to have arisen from chance variations in temperature.

Although this conclusion is logical, one fact suggests that the dip in the curves in late July may represent a true variation. About the middle of July the ice in Barrow Strait, lying south of Resolute, generally breaks up. The open water tends to produce fog and so to reduce the temperatures at Resolute. This relationship shows itself in the mean amounts of low cloud observed at the synoptic observations. These are, for 9 years of data :

22 June to 28 June	50 per cent;	13 July to 19 July	60 per cent;
29 June to 5 July	38 per cent;	20 July to 26 July	60 per cent;
6 July to 12 July	44 per cent;	27 July to 2 Aug.	56 per cent.

Thus although the data are not conclusive, it is quite probable that the mean temperature drops slightly about the middle of July because of the open water that occurs in the vicinity.

3. Time of minimum temperature

The time of minimum temperature is of greater interest because the period is under a different regime than is found in more southerly latitudes, while the influences during the summer months in the Arctic are not greatly different in form, although they do differ in intensity, from those found in temperate locations.

The minimum points of the curves are found on 3 March, with − 27·7° for maximum, and 4 March with − 40·2° for minimum. During the period from 15 December to 4 March the trend is very slowly downward, dropping less than 15° in 11 weeks. This is the trend that should be expected, for as the winter advances the earth cools gradually with the outgoing radiation. During this period the effect of non-cyclical variations becomes more prominent and there are many minor maximum and minimum points on the curves. If the value of 5° is accepted as the maximum correction that may be applied, the time of minimum temperature may be any time between 15 January and 8 March.

By analogy we may compare the temperature trend during this period with the changes during the night in more southerly latitudes. In both cases the sun's heat is cut off, and so the temperature drops rapidly at first. Later cooling causes the balance to approach equilibrium, and the rate of decrease is much slower. Short-term effects, such as brief gusts of wind during the night or cyclonic storms during the winter, cause temporary interruptions in the general trend. The cooling during the night continues until after sunrise, when the radiation from the sun balances the heat loss. After the time of minimum temperature, the warming is rapid. If the analogy is accepted, the minimum temperature in the Arctic should occur two or three weeks after the first appearance of the sun, for there is little heat during the first few days after the sun appears. At Resolute, the maximum solar radiation on a horizontal surface recorded in any day in February was 75 cal cm^{-2}.

4. Study of 6-hr temperature changes

To learn more of the effect of the sun on the temperature, use was made of the observations taken at 00, 06, 12, and 18 hr Central Standard Time (CST). Change of temperature between two observing times is caused by the annual trend, the diurnal trend, and non-cyclical weather changes. The annual trend in temperature may be obtained by means of the curves of Fig. 1, smoothed to eliminate the non-cyclical effects. The diurnal variation during the period when the sun is low is unknown, but it is assumed to be approximately constant for any one period of the day and year. The non-cyclical effects are variable in nature, but it is assumed that the mean of these for a long period for any one time interval is zero. A study was made by 10-day periods of the change $T - T_{h-6}$, where T_h is the temperature for some specific synoptic hour and T_{h-6} the value six hours earlier. The two periods 1 January to 31 March, and 1 September to 30 November, were studied since these cover 5 February, when the sun first makes its appearance after the winter night, and 6 November, when it last appears in the autumn. The data used were the same as for Fig. 1.

TABLE 2. VALUES OF THE MEAN CHANGE OF TEMPERATURE (M) FOR SPECIFIED 6-HR PERIODS, OF TH STANDARD DEVIATION (S.D.) OF THESE CHANGES, AND THE STANDARD ERROR (S.E.) OF THE MEAN CHANGES A RESOLUTE. DATA USED WERE FROM 8 OR 9 YEARS OF OBSERVATIONS BETWEEN OCTOBER 1947 AND MARC 1957 INCLUSIVE. ALL VALUES ARE IN DEGREES FAHRENHEIT. THE LAST COLUMN GIVES THE MEAN 6-HR CHANG CALCULATED FROM THE MEAN ANNUAL TREND

Time (CST)	06 – 00			12 – 06			18 – 12			00 – 18			Mean 6-hr
Period	M	S.D.	S.E.	M	S.D.	S.E.	M	S.D.	S.E.	M	S.D.	S.E.	chang
Sept. 1-10	− 0·55	2·23	0·25	1·61	1·89	0·21	0·12	2·07	0·23	− 1·58	2·07	0·23	− 0·1
Sept. 11-20	0·02	2·61	0·29	1·14	2·33	0·26	− 0·67	2·37	0·27	− 1·10	3·31	0·37	− 0·1
Sept. 21-30	− 0·21	2·82	0·32	1·52	3·52	0·39	− 1·29	2·72	0·30	− 0·48	3·12	0·35	− 0·1
Oct. 2-11	− 0·92	3·73	0·42	1·04	3·94	0·44	0·00	3·48	0·39	− 0·53	3·83	0·43	− 0·1
Oct. 12-21	− 0·79	4·37	0·49	0·15	3·98	0·44	− 0·62	4·16	0·47	− 0·21	5·28	0·59	− 0·1
Oct. 22-31	− 0·01	3·93	0·44	0·29	4·29	0·48	0·15	4·21	0·47	0·06	3·54	0·40	− 0·1
Nov. 1-10	− 0·11	4·19	0·47	0·22	5·05	0·56	− 0·35	3·86	0·43	− 0·48	4·24	0·47	− 0·1
Nov. 11-20	0·08	3·68	0·41	− 0·31	4·33	0·48	− 0·71	4·32	0·48	− 0·40	4·39	0·49	− 0·1
Nov. 21-30	− 0·53	4·14	0·46	0·42	4·98	0·56	0·18	4·09	0·46	0·50	4·28	0·48	− 0·0
Jan. 1-10	− 0·41	4·75	0·50	0·19	4·04	0·43	0·82	3·94	0·42	0·13	4·97	0·52	− 0·0
Jan. 11-20	− 0·95	4·24	0·45	− 0·55	4·23	0·45	− 0·12	4·04	0·43	− 0·40	3·72	0·39	− 0·0
Jan. 21-30	0·25	4·59	0·48	0·26	4·23	0·45	− 0·25	4·21	0·44	− 0·06	5·11	0·54	− 0·0
Jan. 31-Feb. 9	− 0·20	5·01	0·53	− 0·32	4·64	0·49	0·25	4·16	0·44	0·08	4·32	0·46	− 0·0
Feb. 10-19	0·72	5·14	0·54	0·21	4·34	0·46	− 0·33	4·68	0·49	− 0·61	4·34	0·46	− 0·0
Feb. 20-Mar. 1	0·34	5·10	0·54	0·87	3·92	0·41	− 0·19	4·87	0·51	− 1·69	4·07	0·43	0·0
Mar. 2-11	0·24	4·20	0·44	2·49	4·00	0·42	− 0·44	4·10	0·43	− 0·39	4·38	0·46	0·1
Mar. 12-21	− 0·64	5·39	0·57	3·43	4·42	0·47	− 0·54	4·12	0·43	− 2·01	4·74	0·50	0·1
Mar. 22-31	− 0·60	4·67	0·49	4·39	4·53	0·48	− 0·09	4·68	0·49	− 3·66	4·60	0·48	0·1

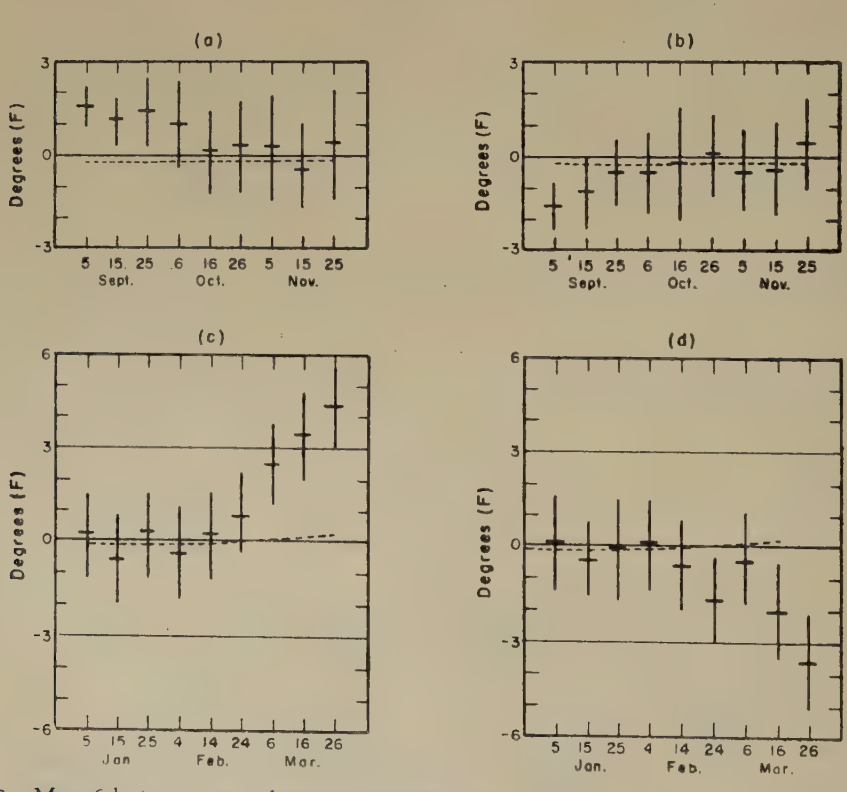

Figure 2. Mean 6-hr temperature changes at Resolute, given by horizontal bars, plus and minus three times their standard errors, given by vertical lines. Dates given are for the mid-points of the 10-day intervals. Dotted lines give the mean 6-hr change computed from the annual trend of temperature. (a) Values of $T_{12} - T_{06}$, September to November; (b) values of $T_{00} - T_{18}$, September to November; (c) values of $T_{12} - T_{06}$, January to March; (d) values of $T_{00} - T_{18}$, January to March.

Table 2 gives for the 18 separate periods the mean temperature changes for the specified six hours, the value of the standard deviation (S.D.) of these values, and also the values of S.D. $(N)^{-\frac{1}{2}}$, where N is the number of observations. This last quantity gives the standard error of the mean. The last column gives the change in temperature for the 6-hr period arising from the annual trend as given in Fig. 1.

As stated above, the mean temperature change is the sum of the changes caused by the annual trend given in the last column, the diurnal trend, and non-cyclical variations. The mean of these latter should be zero for a long series of observations. For a shorter period it may be assumed that the mean change from non-cyclical causes is not greater than three times the standard error of the mean. If the difference between the computed mean and the change arising from the annual trend is greater than three times the standard error, it is assumed that this arises from the diurnal trend of the temperature which, therefore, is significant at that period of the year.

The major changes in temperature from 00 hr to 06 hr, and from 12 hr to 18 hr, during the periods studied are caused by random influences, although theoretically there should be a slight cooling at both periods when the sun is rising and setting. Because the other two periods are more significant in the study of the diurnal trends, the information concerning them has been put into graphical form in Fig. 2. Mean temperature changes are plotted as short horizontal bars. The long vertical lines intersecting these are in length six times the standard errors, thus giving the range of probable values. The mean 6-hr differences given in column 14, Table 2, are shown by broken lines.

According to the values in Table 2 and Fig. 2, the temperature rises from 06 hr to 12 hr during the period 1 September to 30 September are small but significant. The change during the first ten days of October is borderline, but thereafter changes can be attributed to random variations. The temperature drop from 18 hr to 00 hr has little significance after 20 September. In the late winter, the daily rise in temperature from 06 hr to 12 hr is marginal for the period 20 February to 1 March, but significant thereafter. The temperature drop from 18 hr to 00 hr is significant after 11 March, and for the period 20 February to 1 March, and is marginal for the period 2 March to 11 March.

One can then conclude that the effect of the returning sun is noticeable after 20 February although the heating effect may not be great enough to compensate for the radiation from the earth before 1 March. Before the end of the first ten days of March the sun's heat has definitely begun to cause the mean temperature to rise.

The diurnal variation in temperature is established 25 days after the sun first reappears in February, and there is some sign of its influence during the period 15-25 days after 5 February. In contrast, in the autumn the diurnal change has almost disappeared 35 days before, and entirely gone 25 days before 5 November. The contrast is a result of the cloud conditions. February is normally clear and the sun's radiation can reach the earth. October, with open water still in the vicinity, is frequently cloudy, shielding the earth from the weak sun. Its altitude at noon on 20 February is 3°, on 1 March 7°, and on 11 October 8°.

From these results, it is possible to estimate the time of minimum temperature. Obviously it is after 20 February. Since the sun's heat is effective during the period 2 March to 11 March, the time of the reversal of the trend of temperature is prior to the middle of this period and possibly before the beginning. It would appear that the most probable time for the minimum temperature is during the period 25 February to 4 March.

5. Conclusions

According to the results of the foregoing study, the time of maximum temperature at Resolute lies between 7 July and 4 August with the most probable time between 7 July and 15 July. The minimum temperature occurs between 20 February and 5 March, probably between 25 February and 4 March. The 6-hr temperature differences proved to be a valuable tool in assisting location of the period when the influence of the sun was absent from the station.

6. Acknowledgments

The author takes this opportunity to acknowledge the helpful suggestions made by his colleagues, Mr. M. K. Thomas and Mr. D. W. Boyd, in the preparation of this paper. The paper is published with the permission of the Director, Meteorological Branch, Department of Transport, Canada.

Reference

Markham, A. H. 1884 *The great frozen sea* (Kegan Paul, Trench), p. 222.

Correspondence

551.461.2 : 551.515.8

A preliminary investigation of long waves at Newlyn

By J. Darbyshire

There is an arithmetical error in p. 69, line 10, of the above paper, published in the January 1958 issue.

for gives $T = 165$ min . . . . The velocity c is 87 ft/sec or 52 kt

read gives $T = 27$ min . . . . The velocity c is 59 ft/sec or 35 kt

This shows that the long waves could have been edge waves, although, as stated in the paper the evidence is not very conclusive.

It is proposed in the near future to prepare an electronic analogue model of this bay and this may throw further light on the subject.

National Institute of Oceanography,
Wormley,
Godalming, Surrey.
1 April 1958.

551.558.21

Airflow over mountains: indeterminacy of solution

By Enok Palm

In a recent number of this *Journal* (*Q.J.*, **84**, p. 182, 1958) R. S. Scorer has written an article on two-dimensional airflow over mountains. Some of the assertions in this article deserve comment.

Intending to apply the Fourier theorem, Scorer considers, as is customary, first the flow over corrugations

$$\zeta_1 = \cos kx \qquad . \quad . \quad . \quad . \quad . \quad . \quad (1)$$

The differential equation for ζ, the vertical displacement of the particle at (x, z) (disregarding the kinematical effect of the density variation) will be of the form

$$\frac{d^2\zeta}{dz^2} + (l^2 - k^2)\,\zeta = 0 \qquad . \quad . \quad . \quad . \quad . \quad (2)$$

and if l is independent of z, this has solutions

$$\zeta = A \cos \mu z + B \sin \mu z \qquad . \quad . \quad . \quad . \quad (3)$$

where

$$\mu = (l^2 - k^2)^{\frac{1}{2}} \qquad . \quad . \quad . \quad . \quad . \quad . \quad (4)$$

When $l < k$, A and B are determined uniquely by Eq. (1) and by the requirement of finite (i.e., not infinite) value of ζ at infinity, the solutions becoming exponential in type. When $l > k$, the second requirement is fulfilled for arbitrary values of A and B, the first condition determines only A, and the solution seems to be incompletely determined.

Following Scorer the solution may in this case be written

$$\zeta = \cos kx \cos \mu z + B \cos kx \sin \mu z \qquad . \quad . \quad . \quad (5)$$

in which B is *any* function of k. Quoting from Scorer '*In the absence of a special second condition waves corresponding to no disturbance at the ground should be excluded for the same reason that lee waves on the upstream side of the mountain are excluded.*' (Italics by Scorer). The solution is then

$$\zeta = \cos kx \cos \mu z \qquad . \quad . \quad . \quad . \quad . \quad . \quad (6)$$

This assertion is, as will be shown below, incorrect.

1. What is observed in nature is that the waves are created during some initial interval and approach a (quasi) stationary form. In other words, the problem has to be treated as an initial-

value problem. This was first pointed out by Höiland. Wurtele (1953) and Palm (1953), independently, proved that for increasing values of time the solution approaches a limit. This limit is not the form given by Scorer (Eq. 6), but the solution earlier found by Lyra (1943) and Queney (1947) by using the device of introducing an artificial friction, namely

$$\zeta = \cos(kx + wz) \quad . \quad . \quad . \quad . \quad . \quad . \quad (7)$$

2. The indeterminacy of the solution may also be eliminated by applying the radiation condition, i.e., by the requirement that the energy flow in a frame of reference where the mountain is moving, is directed away from the mountain, as shown by Eliassen and Palm (1954). The result is identical with Eq. (7).

3. In the case of lee waves, i.e., when a discrete set of resonance waves exists, the waves on the upstream side of the mountain are excluded for instance by considering the problem as an initial-value problem. *Therefore, in the absence of a special second condition (such as required by a rigid lid) the correct solution in the actual problem is that given by Eq.* (7), *and not that of Eq.* (6), *for the same reason that lee waves on the upstream side of the mountain are excluded.*

This result may also be proved directly in the following way. Assume for the moment

$$l = ae^{-bz} \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (8)$$

If b is small, the solution for an arbitrarily-shaped mountain will consist of a number of lee waves plus an integral necessary to make the solution non-singular and to satisfy the boundary condition at $z = 0$. When $b \to 0$, l tends towards a constant; and the number of lee waves may be shown to tend towards a continuous infinity. The solution approaches a limit, and this limit corresponds to Eq. (7).

A similar way of reasoning was given by Corby and Sawyer (1958). They introduced into the problem a horizontal rigid lid which was displaced to infinity. They also found Eq. (7).

It may be of interest to point out that considering in this problem (l constant) waves propagating vertically, it is found that waves with a phase velocity downwards have an energy flow (group velocity) upwards and vice-versa (Eliassen and Palm 1954). From this it follows that the radiation condition mentioned above is not identical analytically with Sommerfield's radiation condition since he discussed waves with group velocity and phase velocity in the same direction. The difference will appear in a change of sign.

What is pointed out above is also valid when the model consists of two or more layers.

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Wurtele, M.	1953	*Univ. Calif., Met. Dept. Sci. Rep., No.* 3, AF 19 (122)-263.

Institute for Weather and Climate Research,
Astrofysisk Institutt,
Blindern, Oslo,
Norway.
9 June 1958.

Reply

By R. S. Scorer

Corby and Sawyer (*Q.J.*, **84**, p. 284, 1958) and Palm have failed to see the point of my letter. I have not argued that their solutions, arrived at by various devices, are wrong.

First it should be made clear that the fact that the same solution is arrived at by two or more different devices does not mean that the flow represented by it will always occur. The lid of Corby and Sawyer which is removed very slowly to infinity while the intervening space is filled with fluid behaving in the correct manner is not found in nature. Nor does the flow necessarily start up from rest in the manner discussed separately by Höiland, Wurtele, and Palm.

Nor is the nature of the flow determined by friction as is assumed in Lyra's treatment and in Queney's early papers – as indeed Queney himself has shown, because the time required for friction to have a significant effect is too long, the particles passing through the region under consideration in a much smaller time.

When a suitable boundary condition is imposed a solution is arrived at. My italicized remarks, quoted by Palm, deal with the case in which there is no such condition; and they are not proved incorrect by considering conditions such as those discussed by Palm. If the effect of friction is discounted the train of lee-type waves which can, from the mathematical viewpoint, occur on the upstream side of a hill, can be excluded on the ground that they do not correspond to anything producing them. They can be added or subtracted quite freely and still a solution of the problem is obtained. The particular solution to choose is the one which can be attributed to the mountain. By the same argument, *when there is an infinity of solutions* the particular one chosen should omit all those which are not attributable to the mountain. Of course if Corby, Sawyer and Palm exclude an infinity of solutions by imposing a condition there is no more to be said about the solution they arrive at except that it applies under the condition stipulated. In nature these particular stipulated conditions do not in fact occur.

The waves may begin in a variety of ways. For instance, there may be flow with no waves, with eddies in the lee, and the wind or stability profiles change so that waves do occur. Palm states that ' what is observed in nature is that the waves are created during some initial interval and approach a stationary form.' But this does not mean that the flow starts from rest. Far from it ! What about ' evening waves ' ? Since there is an infinite number of ways in which waves can happen it is probable that any solution which is possible can be arrived at by giving the system a suitable history; i.e., by imposing suitable boundary or initial conditions. It is not correct to assume that a solution arrived at by a device, particularly if it involves letting a parameter tend to zero or infinity, is the one which nature would produce if the parameter were always at the limit. For instance inviscid flow is not the same as flow at very large Reynolds numbers.

Thus my object was not to pose a particular (and probably unrealistic) problem in order to eliminate the indeterminacy so that it could thereafter be forgotten about, but to see what could be said when the solution was mathematically indeterminate.

With regard to the flow of energy, I showed (*Q.J.*, **80**, 1954, p. 419) that the rate of working by the lower layers on the upper was exactly cancelled by the downward transport of kinetic energy for any waves, so that there is no flow of energy. But in any case there is no valid reason for supposing that energy should flow upwards rather than downwards. The work must come from flow across the isobars and this may occur at any level, so that if there were a flow of energy it could be downwards and the airstream could suck the mountain towards the direction from which it flowed !

Then, can a streamline be frozen and a correct solution arrived at ? If we take the case of flow over a corrugated ground of form $\cos kx$ under a rigid lid, in certain airstreams there are several nodal surfaces. If we apply the principle that any streamline may be frozen and a correct solution obtained we may freeze one of the nodal surfaces and conclude that the flow between two rigid boundaries is of wavelength $2\pi/k$ and of some particular amplitude ! Undoubtedly a correct solution is arrived at by freezing a streamline but since the problem is indeterminate it is not the unique solution. Corby and Sawyer argue that there cannot be nodal surfaces if there is a whole spectrum of wavelengths present but this is irrelevant to the question of whether a streamline can be frozen; all it means is that in that case the procedure cannot be made to look quite so ridiculous.

Finally, with regard to the mathematics of the general solution. I was concerned to point out that in the general solution

$$\zeta = A \cos \mu z + B \sin \mu z$$

the term B could lead to something quite arbitrary. This arbitrariness is removed by the various devices of Corby, Sawyer, Palm and others. But nature does not necessarily employ such simple devices, particularly when the flow varies with time in a complicated way. What if nature does not bother to employ any device at all ? Then according to my argument $B = 0$.

Department of Mathematics,
Imperial College,
London S.W.7.
22 July 1958.

551.509.317

An essay on low-latitude wind analysis

Reply

By J. R. Scott

By omitting the words ' particular type of ' from the sentence under criticism I have strayed from the path of accurate reporting, and offer an apology to Professor C. E. Palmer for this omission. I deal with the two ' errors ' he has detected in the following paragraphs.

Streamline confluence or ' convergence ' is a scalar quantity whose distribution is relatively objectively obtainable from an isogon diagram. It is representable by isopleths of ' rate of change of direction along the normal to direction of flow.' I have constructed a number of such charts using data which included most of the sets published by Palmer (*Q.J.*, **78**, 1952). In no case was there suggestion that streamline confluence was heavily concentrated along extensive lines. For this reason I found the emphasis placed by Palmer and others on ' asymptotes of convergence ' puzzling. Most meteorological textbooks consider actual convergence as occurring along geometrical lines or at geometrical points only, when discussing linear fields of flow and their associated streamlines. I could only ascribe this emphasis as being due to carrying this limitation into their ideas of the characteristics of non-linear fields of flow. This I regard, as far as low-latitude fields are concerned, as a preconception which cannot be justified, even with the dense networks of winds used by Palmer. I am therefore now surprised to hear of a machine which can construct streamlines which embody this preconceived notion. In view of my own results I remain sceptical that the final product of this machine is as objective as Professor Palmer claims. These considerations apart, machines are not in general use. From the sources of information available to me it seems that the general practice is to insert streamlines without even constructing the isogons.

With regard to the second ' error ' the original statement I criticised was that the type of cloud line concerned *always* coincided with a streamline, and I pointed out that under this condition it could not move transversely. Professor Palmer interprets this as meaning that I have implied that an ' asymptote ' cannot move parallel to itself (I assume that he means ' normal to its own direction ' here). There is indeed confusion here. I will confine myself to a justification of my own statement. Weather systems which move as entities and the wind field at a single level are the concern here. Therefore development and decay of weather and wind shear in the vertical are not considered. Consider a zone of convergence which moves horizontally as an entity. At each instant it may be regarded as being fed by a purely convergent system of winds which is representable by isopleths as described in my paper. In whatever manner these isopleths may vary with time they must retain the characteristic of indicating zero wind in the convergence zone itself. The remaining part of the wind field comprises similar wind systems feeding other convergence areas and the solenoidal part. The only possibility of the zone moving as an entity is that this remaining part translates it, and this it can only do with its own velocity. Thus a line of convergence which remains an entity can only move transversely if there is a component of *actual* wind in it normal to it. If the convergence line *always* coincides with a streamline such a component never exists, and it follows that transverse movement does not occur. The movement of temperate latitude fronts provides good illustration of these principles.

If on every occasion of wind analysis a convergence line which has moved as an entity is found to coincide with a streamline, I would prefer to suspect the method of wind analysis rather than claim an empirical discovery involving the ineffectiveness of convergence unless associated with an entity found from wind directions alone.

Meteorological Office,
P.O. Box 715,
Singapore.
7 June 1958.

551.515.8 : 551.557.3 : 551.589.1

A synoptic study of West African disturbance lines

By R. H. ELDRIDGE

(Read 19 March 1958. See *Q.J.*, **83**, p. 303)

DR. A. G. FORSDYKE : One of the greatest difficulties in the study of tropical meteorology is that of classifying synoptic models, which arises on account both of their indefinite nature and of the sparse observational network. Mr. Eldridge is to be congratulated on an analysis which puts some sort of order into West African disturbances.

To my mind, the most intriguing thing about these disturbances is their linear extension. Mr. Eldridge suggests that this may be explained by the undercutting of warm moist unstable air by the precipitation-cooled air; the undercutting takes place in all directions except that opposed to the general air flow, so that the disturbance extends longitudinally and advances on a broad front. I consider that it is necessary to invoke some more over-riding cause, and suggest a perturbation travelling in the general east-west current, of a similar nature to, but less pronounced than, the Easterly Wave. The more violent weather associated with the D/L than with the Easterly Wave would then be explained by instability and the juxtaposition of air masses with widely differing properties.

The available observations do not, of course, afford strong evidence for the linear extension of D/Ls. I am therefore particularly gratified to see Mr. Eldridge's slides of radar echoes from D/Ls over the sea (Figs. 1 and 2 below, not reproduced in the paper), which afford striking evidence of their linear extension and broadside progression.

MR. J. S. SAWYER : The dynamical structure and mechanism of West African disturbance lines has always seemed to me to be different from that of the Easterly Waves found in oceanic regions of the Equatorial Belt. As Mr. Eldridge suggests, density differences arising from the cooling of air by rain probably play an essential part in the D/L but not in the Easterly Wave.

MR. D. J. SCHOVE : I was particularly interested in Mr. Eldridge's paper as I worked in West Africa some years ago. I should like to know whether the upper-air data now available enable Mr. Eldridge to comment on the structure of a D/L which I put forward (*Q.J.*, **72**, p. 105) and if he has found any relation between extra-tropical disturbances and the start of a D/L. Also, has he been able to relate any D/Ls to previous disturbances in East Africa, or subsequent ones in the Caribbean area or the northern part of South America ?

DR. H. L. PENMAN : Is it possible that, given an initial easterly movement, a disturbance such as is described in the paper could continue in this direction ?

MR. P. BALL : Would Mr. Eldridge say whether the movement of the D/Ls was slowed down by local configuration, such as a range of hills, or even an isolated, but perhaps significant contour change ?

MR. R. H. ELDRIDGE (*in reply*) : In reply to Dr. Forsdyke, it seems to me *a priori* preferable, when considering the origin of a tropical weather phenomenon, to look for one related as closely as possible to the energy of incoming solar radiation. This I have tried to do in Section 9 of the paper. I still feel that this explanation of the development of a line is plausible and fits in with all the available information, inadequately worked out as it inevitably is. The postulation of a ' travelling perturbation ' seems rather less satisfactory and to bring in further problems.

I am afraid that the lack of upper-air data and the lack of synoptic information over the sea precludes replies to several of Mr. Schove's points. One set of conditions favourable for the formation of D/Ls is considered in Section 4 (c). I am not satisfied that any others can yet be substantiated. I think there is little doubt that a few D/Ls which affect Africa west of 10°E may start east of 25°E but I was not able to find an indisputable case of this in 1955. It is still all too easy to carry over concepts from temperate-latitude meteorology, hence our use of the non-committal colour brown for the position of a D/L and the avoidance of all terms generally associated with polar fronts.

Two comments are relevant to Dr. Penman's point. Firstly, in 1955, over the area considered in the paper, I could find no D/L type of phenomenon which moved with a component from west to east. Secondly, whether or not the suggestion is possible, the known kinematic data are against it. For while some D/Ls appear to move through the atmosphere at a speed of the order of 15 kt, upper-wind statistics show that in the D/L seasons a substantial depth of the atmosphere usually has a westward component of this order or greater.

The synoptic network is not sufficiently fine to give a full answer to Mr. Ball's question, but it seems that even such a range as the Togoland hills does not have a marked slowing-down

effect. But hills may set off thunderstorms before the arrival of a D/L and this has, for example, in the afternoon or evening, the effect of increasing locally the speed of a D/L. This was certainly visible at times over the Aburi scarp to the north of Accra.

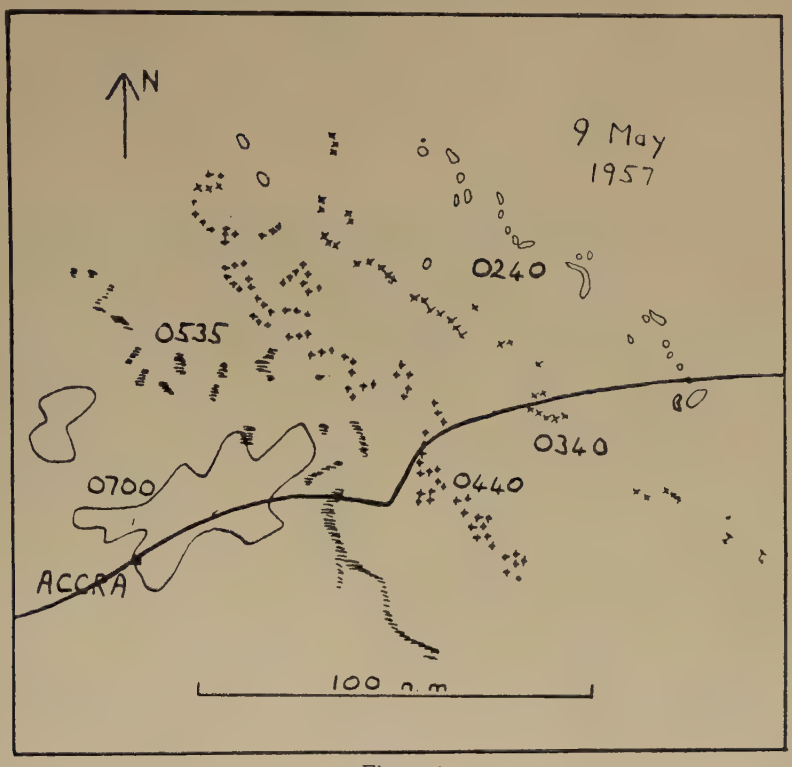

Figure 1

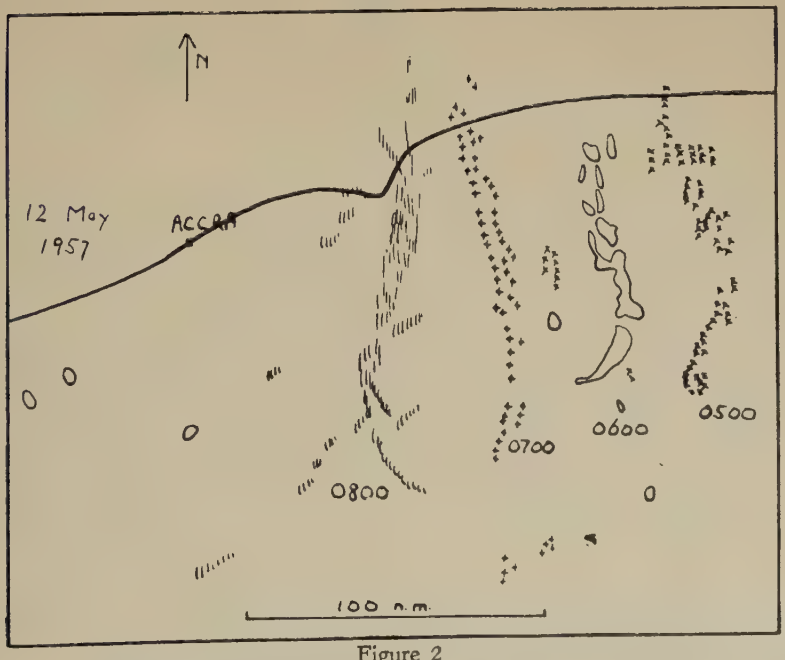

Figure 2

Figures 1 and 2. Successive positions of D/Ls of 9 and 12 May 1957, shown on radar screen at Accra. Times are GMT. When presenting his paper, Mr. Eldridge showed the above two diagrams. These give successive positions of two D/Ls as shown on the screen (PPI) of a Decca Type-41 weather radar (3 cm) at Accra.

551.557.31 : 551.511.

Theoretical studies of diurnal wind-structure variations in the planetary boundary layer

By K. BUAJITTI and A. K. BLACKADAR

(Read 19 March 1958. Presented by Dr. D. R. Davies. See *Q.J.*, **83**, p. 486)

Mr. E. Gold, F.R.S. : I think the main value of this paper lies in the observational results. Any trustworthy observations of the actual diurnal variation of wind at different heights is illuminating. In this case the large increase in the amplitude of the variation up to the 2,650 ft level is surprising. In 'periodic variations of pressure, temperature and wind in the atmosphere,' *Phil. Mag.*, Jan. 1910, I concluded that the diurnal variation of pressure decreased upwards and vanished at about 4,000 ft. Above that level it reappears with a phase difference of 180° and increases to $1\frac{1}{2}$ times the surface value at about 20,000 ft. The corresponding diurnal variation of wind would, with this system, vanish at about 4,000 ft and then increase to about 3 times the 'surface' value. Clearly the conditions over Oklahoma in summer do not conform to this system.

Mr. J. S. Sawyer : Discussion of the wind distribution in the friction layer has usually been on the basis of steady flow. I was therefore interested to see that the authors retain the inertia terms $\partial u^*/\partial t$, $\partial v^*/\partial t$ in their Eq. (3). Consideration of the magnitude of the terms in the equation shows that the wind must take several hours to adjust itself to a change in the turbulence regime, but I had not seen this brought out previously. This has the significance that with westerly winds in the British Isles the wind will often hardly have time to adjust its profile according to the change in surface roughness from sea to land before the air has passed across the country.

Dr. D. R. Davies (*communicated*) : The recorded observational results are of considerable importance, but the interpretation of these results is, I think, also significant. Any observational work and theoretical analysis which may throw some light on the problem of understanding the structure of atmospheric turbulence, on any scale, is very welcome. In numerical forecasting, for example, we surely need to know what mathematical forms to use for the turbulence terms in the equations of motion. In some meteorological situations perhaps the inertia terms dominate, as assumed at present, but in other cases neglect of the Reynolds stresses may well upset the calculation. Further work of this kind in other conditions may promote a solution of this problem.

It would have been helpful if the authors had given us more details of the meteorological conditions prevailing during their observational work. The development of a direct method of integrating the equations of mean flow, replacing the authors' trial and error method, would also be helpful. An extension of the theoretical analysis to include thermal wind effects would be a considerable advance.

Dr. A. K. Blackadar (*in reply, communicated*) : Mr. Gold treated friction as a force which is opposed to the velocity and proportional in magnitude. Such a treatment does not take into account the large diurnal variation of exchange coefficient, and so does not explain the very large wind variations which occur in continental locations. I believe that the theoretical analysis is of value in demonstrating that this mechanism of friction is the dominant cause of the observed variations in this region rather than other mechanisms which have been suggested.

Most of the observations on which Fig. 1 was based were made under conditions of southerly or southwesterly flow. The skies are mostly clear, giving rise to unstable lapse rates in the daytime and strong inversions up to between 1,000 and 1,500 ft at night. Recent studies of typical conditions in this area indicate a decrease of geostrophic wind with height resulting from higher temperature to the left of the flow. It can easily be shown that when the exchange coefficient is independent of height, the characteristics of the oscillation are not affected so long as the geostrophic wind varies linearly. Under more general conditions it is not obvious what result this condition has on the solutions, but it would not be a difficult matter to run such a problem on a computer with a greater capacity than that which we used.

A direct method of integrating the equations has been developed by Mr. K. Ooyama as his M.S. dissertation at New York University, and was used to verify the best solution of the computer. However, this method is extremely laborious, and therefore is not so useful for experimenting. I have also attempted to evaluate the distribution of exchange coefficient from the primitive (un-integrated) equations during a different period but so far the results have been inconsistent, probably because of the uncertainty of geostrophic wind.

551.551.25

The measurement of gustiness in the first few thousand feet of the atmosphere

By J. I. P. JONES and H. E. BUTLER

(Read 18 June 1958. See *Q.J.*, **84**, p. 17)

Mr. B. C. V. Oddie : I would like to remark on the exceptional interest of the records displayed by Mr. Jones. One in particular includes a cross-section of a small cumulus, and shows a wide region of descending air with very little turbulence ahead of the cloud and ascent and considerable turbulence over the whole width of the cloud, both within and below it. Much of this detail has previously been inferred, but never so clearly shown : more records of this kind are very desirable.

Mr. J. I. P. Jones (*in reply on behalf of the authors*) : I would like to say that we are continuing to make records of vertical gustiness from time to time in connection with our programme on vertical diffusion, and it is hoped that the accumulated data will appear in due course.

551.507.321.2 : 551.510.42 : 551.551.8

The use of balloons in atmospheric pollution research

By D. H. LUCAS, G. SPURR and F. WILLIAMS

(Read 18 June 1958. See *Q.J.*, **83**, p. 508)

Mr. B. C. V. Oddie : I think it remarkable that the balloon stayed in the plume, which on visual evidence mixed with the surrounding air (and therefore cooled) very rapidly indeed. The gas in the balloon could not cool so fast, so that the balloon should surely have some buoyancy to carry it out of the smoke.

Dr. R. S. Scorer : Measurements made on thermals at Imperial College show that particles cannot escape from the top unless their free ascent velocity is comparable with the rate of rise of the thermal because the fluid in the path of a thermal is mixed into it.

At the base some of the exterior air is entrained up into the centre in such a way that particles with a free fall speed of less than $0{\cdot}4\ w_c$, where w_c is the rate of rise of the cap, cannot escape. Only particles with a fall speed in excess of $1{\cdot}6\ w_c$ necessarily escape. Particles with intermediate fall speeds may or may not fall out (this work is soon to be reported upon by Miss Woodward).

The corresponding figures for a bent-over plume are likely to be rather similar ; consequently Mr. Spurr's balloons are likely to remain in the plume for a considerable distance even if they have a speed of ascent or sink relative to the plume air.

Mr. G. Spurr (*in reply on behalf of the authors*) : Dr. Scorer has to a large extent answered Mr. Oddie's query. As could be seen clearly in the demonstration film, the balloons remain in the plume almost as though they were held by the plume. I agree that the balloon will lose heat less quickly than the plume on emerging from the chimney, but the balloon is unlikely to reach the temperature of the gas during its passage up the chimney. These two factors appear to cancel each other out.

551.510.411

The hydrogen content of atmospheric air at ground level

By E. GLUECKAUF and G. P. KITT

(Read 18 June 1958. See *Q.J.*, **83**, p. 522)

Prof. P. A. SHEPPARD (*President*) : The meteorological interest of such work as the authors have reported partly depends on having some ideas about the nature of the hydrogen cycle in the atmosphere. One realises also that ideas about the cycle must spring partly from taking observations, and that the early observations are likely to be most productive of such ideas. Have the authors, either from theory or their own previous observations, formed any rough views about the dominant factors in the hydrogen cycle ?

Dr. F. J. SCRASE : I wonder if anyone has computed how much hydrogen is put into the air by Meteorological Services ?

Dr. R. S. SCORER : If the source of hydrogen is in the upper atmosphere – which I doubt because it is a region of escape and is a source of oxygen – no doubt ground-level concentrations would be greatest in February. But equally if there were any terrestrial – particularly industrial – source, concentrations due to it would also be greater in February than in August, just as for any other terrestrial source of pollution.

Dr. E. GLUECKAUF (*in reply to the President*) : I feel that it is most likely that the hydrogen in the atmosphere is of terrestrial origin, to which organic, volcanic and industrial activities may contribute. Like Dr. Scorer, I would expect that the upper atmosphere with its free oxygen atoms and its rate of hydrogen escape must act as a sink for molecular hydrogen, and consequently have a lower H_2 content than the troposphere. If the observed fluctuations in hydrogen content are real, I should try to interpret them as a consequence of vertical mixing, produced by large air masses from the upper atmosphere entering the tropospheric air currents.

In reply to Dr. Scrase, the hydrogen content of the atmosphere is of the order of 2,000 km^3 N.T.P. The amounts used for ammonia manufacture may be of the order of 10 cb-km per year, while the meteorologically-used quantities are no doubt insignificant compared with either volume.

551.521.3 : 551.521.61

A direct measurement of infra-red radiation divergence to 160 mb

By V. E. SUOMI, D. O. STALEY and P. M. KUHN

(Read 16 June 1958. See *Q.J.*, **84**, p. 134)

DR. J. T. HOUGHTON : I am sure that Dr. Staley and his collaborators have provided us with an instrument which will be of considerable use in the lower atmosphere. I would like, however, to have more information about the accuracy of the instrument at high levels in the atmosphere, where the corrections to the instrument are much larger than the radiation flux it is required to measure. For instance, at the tropopause the downward radiation energy is about 3 milliwatts/cm^2. This is measured by the radiometer as the difference between the radiative emission of the element of about 10 mw/cm^2 and the conduction of heat into the element from above and below of about 7 mw/cm^2. Thus, in order to measure the downward radiative energy of 3 mw/cm^2 with a maximum error of 0·1 mw/cm^2 it is necessary to measure the temperature of the element to better than 0·5°C and the conduction terms to better than 1 per cent. To calculate heating or cooling rates due to radiative divergence to better than 0·1°C/day, it is necessary to measure the radiation streams over layers 100 metres thick to better than 0·1 mw/cm^2.

I would also like to know what absorbing-black has been used in the construction of the radiometer, and what data are available on its absorption at long wavelengths. So far as I know, the only material of high absorption at wavelengths beyond 30 μ whose properties have been measured at these wavelengths, is evaporated gold-black. Since most of the divergence of radiation around the tropopause is probably due to the long-wave band of water-vapour at wavelengths beyond 30 μ, it is most important that instruments used for measuring radiation around the tropopause be covered with a black of high absorption at these long wavelengths.

Dr. G. D. Robinson : The angular distribution of the upward and downward streams of radiation must be markedly different. (The upward stream would be expected to be greatest, and the downward stream least, normal to the receiver surface). The sensitivity of the receiver seems likely to vary considerably with angle of incidence. Are these circumstances adequately covered in calibration, and does Dr. Staley think they could seriously affect the readings at great heights ?

Mr. D. D. Clark : What is the effect on the reading of the radiometer of the swinging of the sonde at the end of its connecting cable ?

Dr. Staley (*in reply on behalf of the authors, communicated*) : Dr. Houghton raises an important point. Certainly the downward flux must ultimately tend to zero as one goes to higher and higher levels, and in the upper troposphere downward flux is measured as the relatively small difference of two terms of similar magnitude but the same sign. We must agree with Dr. Houghton's estimates of magnitudes of radiation and conduction terms and the consequences of a maximum temperature error of 0·5°C. In a typical ascent, 15 April 1958, at 20 mb we find 9·74 mw/cm^2 for the radiation term and 7·39 mw/cm^2 for the sum of the conduction terms, resulting in a downward flux of 2·35 mw/cm^2. Had the measurement of the temperature of the upward-facing radiometer surface been in error by 0·5°C, the resulting error in the downward flux would have been 0·24 mw/cm^2, or roughly 10 per cent of the downward flux itself.

Temperature errors can be of two sorts, random or systematic. Random errors appear to be less than 0·1°C, for successive temperature measurements form a smooth progression of points, and lead to regular rather than erratic radiative cooling curves when layers of 20 to 30 mb are employed. Probable systematic temperature errors have been evaluated by Kochanski and Johannessen (Air Weather Service Technical Report 105-108, May 1953) at 0·5°C. We feel such temperature errors may often be much less, especially since our instrument is suspended 100 ft. below the balloon and our ground record and sensors are carefully calibrated. Top, bottom and air sensors are individually calibrated in the circuit of the radiosonde.

Of more importance for the accuracy of the downward flux at high levels is the error term, E_t, which, in its present inclusiveness, we estimate to be no larger than 10 per cent of the largest terms. Near the tropopause and above, it may be 25 per cent of the difference of the largest terms, which gives an error of 25 per cent in the measurement of the downward flux. This estimation makes use of the numerical values cited in the evaluation above.

The growth with height of percentage error in the downward flux must depend on the rapidity of decrease of downward flux. We may cite the results of three typical ascents in the spring of 1958 which indicate, at worst, a very slow decrease with height of the downward flux from the tropopause to the average upper limit of 25 mb reached by our radiosonde flights :

Date	Tropopause height (mb)	$R_{1\downarrow}$ (mw/cm^2)	Burst-point (mb)	$R_{1\downarrow}$ (mw/cm^2)	Weather
26 Mar. 58	240	4·9	50	3·5	Clear
11 Apr. 58	250	7·0	75	4·9	Clear
15 June 58	205	7·0	20	4·2	Clear

Downward flux must to some extent follow the temperature, which ceases to fall and may even increase again from the tropopause to the burst-point.

In the initial stages of our investigation the measurements did not reach the higher levels, and we were primarily interested in the net radiation. For this purpose the error terms, E_t and E_b, are subtracted and tend to cancel, so that the error in the downward flux is unimportant. All recent measurements have reached much higher levels and we are attempting to improve the accuracy of the downward flux. Lateral heat conduction and polyethylene emission can be taken into account and stray radiation eliminated. We feel that we can keep the error in the downward flux measurement under 10 per cent.

Heating or cooling rates are determined by the divergence of the net radiation. In obtaining the net radiation, several of the sources of error cancel, giving an error which is less than 10 per cent. Lateral heat conduction may be taken into account to reduce the error even further. Whatever systematic error exists in the temperature measurement will also cancel, since only the change of temperature difference with height determines the cooling. Thus the individual radiative fluxes need not be measured to better than 0·1 mw/cm^2.

In answer to Dr. Houghton's second question, we have used the W. P. Fuller Company, Los Angeles, California, Velvet Black Decoret Enamel as the blackening agent on the two sensing surfaces of the radiometer. Its spectral characteristics to 40 μ have been determined by the infra-red spectrometry division of the University of Wisconsin Chemistry Department and are reflected in our constant, K. Absorptivity decreases from 0·855 at 30 μ to 0·845 at 40 μ, and shows no tendency to drop off more rapidly with increasing wavelength.

Dr. Robinson asks about the dependence of absorptivity on angle of incidence. The absorptivities used in the calibration were determined for zenith angles up to 45°, but absorptivity showed no dependence on angle of incidence. Accordingly we expect no effect on the results at any height.

Mr. Clark asks about the effect of swinging of the radiosonde. Swing of the radiosonde at ground level through zenith angles up to 45° failed to produce a perceptible difference in the sensor readings. When the instrument was held at 45°, sensor temperatures gradually changed by less than 0·5C. Moreover, we observed the period and amplitude of the swing in flight by means of a telescope and found them to be 5 sec and 25° from the zenith.

Proceedings at meetings of the Society

Special General Meeting, 18 June 1958

A Special General Meeting of Fellows was held to elect the Officers and Council for the year 1958-59.

Fellows had been circulated on 10 April with the list of nominations by Council for Officers and Council for the year 1958-59. No other nominations having since been received, the persons so nominated were declared to be elected unopposed in accordance with By-law 51.

The Officers and Council for 1958-59 were therefore elected as follows :

PRESIDENT

Professor PERCIVAL ALBERT SHEPPARD, B.Sc., F.Inst.P.

VICE-PRESIDENTS

ALAN W. BREWER, Ph.D.
HOWARD LATIMER PENMAN, M.Sc., Ph.D., F.Inst.P.
CHARLES HENRY BRIAN PRIESTLEY, M.A., Sc.D., F.Inst.P.
REGINALD COCKCROFT SUTCLIFFE, O.B.E., Ph.D., F.R.S.

VICE-PRESIDENT FOR CANADA

ROBERT CAMPBELL GRAHAM, M.A.,

VICE-PRESIDENT FOR SCOTLAND

JAMES ROBERT ATKINSON, M.A.

TREASURER

BRIAN CECIL VERNON ODDIE, B.Sc.

SECRETARIES

PATRICK JOHN MEADE, O.B.E., B.Sc.
RICHARD SEGAR SCORER, M.A., Ph.D.
JOHN STANLEY SAWYER, M.A. (Editor)
ARTHUR GEORGE FORSDYKE, Ph.D., A.R.C.S. (Librarian)

COUNCILLORS

GEORGE ARTHUR CORBY, B.Sc.
Instr. Capt. CHARLES ROY DARLINGTON, B.Sc., R.N.
THOMAS HERBERT ELLISON, M.A., Ph.D.
FRANK HENRY WINN GREEN, M.Sc.
ROBERT FREDERICK JONES, B.A.
BASIL JOHN MASON, D.Sc.
ROBERT JAMES MURGATROYD, Ph.D.
EDWARD THOMAS PIERCE, Ph.D.
REGINALD CHARLES RAINEY, Ph.D.
JOHN BRODRICK RIGG, M.A.
CHARLES DESMOND WALSHAW, M.A., Ph.D.
HENRY LIONEL WRIGHT, M.A.

Ordinary Meeting, 18 June 1958

The following candidates were declared to have been elected that day by the Council :

FELLOWS

ABRAM, Robert James, 25 Ettrick Avenue, Fleetwood, Lancs.
CRIPPS, Kenneth Major, B.Sc., 6 Katsonis Street, Nicosia, Cyprus.
DENNIS, Frederick Philip, 4 Beech Way, Twickenham, Middx.
EMERY, Paul Frederick, 67 Boxtree Road, Harrow Weald, Middx.
FOUNTAIN, Leslie John, B.O.A.C., Airways House, 59 Front Street, Hamilton, Bermuda.
GAN, Tong Liang, B.Sc., 764 Mountbatten Road, Singapore 15, Malaya.
HARRIS, Flt. Lt. Frederick Richard, 3 Parade Passage, Penzance, Cornwall.
NAIR, Pattazhy Kerulan Rajendran, Indira Bhavan, Kottarakara, Kearla, India.
SAVIDGE, John P., B.Sc., ' Milwood,' Spital, Wirral, Cheshire (*transfer from Student Membership*).
WHITNEY, Joseph Bevan Robertson, M.A., Ower Quay, Corfe Castle, Dorset.
WINTERS, Richard, 64 Victoria Street, Melton Mowbray, Leics.

FOREIGN MEMBERS

KAUFMAN, Paul, Speer Str. 5, Zurich/Kloten, Switzerland.
WIGGAN, Bernard, L., A.B., U.S. Weather Bureau, Buffalo, New York, U.S.A.

STUDENT MEMBER

HARRISON, Frank Robin, 55 The Highway, Hawarden, Chester.

The following papers were read and discussed :

1. ' The measurement of gustiness in the first few thousand feet of the atmosphere ' by J. I. P. Jones and H. E. Butler (*Q.J.*, **84**, p. 517).
2. ' The use of balloons in atmospheric pollution research ' by D. H. Lucas, G. Spurr and F. Williams (*Q.J.*, **83**, p. 508).
3. ' A simple grating spectrometer for sky-emission studies ' by R. M. Goody (*Q.J.*, **83**, p. 517).
4. ' The hydrogen content of atmospheric air at ground level ' by E. Glueckauf and G. P. Kitt (*Q.J.*, **83**, p. 522).
5. ' A direct measurement of infra-red radiation divergence to 160 mb ' by V. E. Suomi, D. A. Staley and P. M. Kuhn (*Q.J.*, **84**, p. 134).

Summer Meeting, Dublin, 25-27 June 1958

The fifth two-day Summer Meeting of the Society was held in Dublin by kind invitation of the Department of Meteorology and Geophysics, School of Cosmic Physics, Dublin, and took place in the Museum Buildings, Trinity College by permission of the Provost.

The meeting opened on the evening of 25 June with a reception given by the Provost of Trinity College, Dublin.

Two symposia were held. The President, Professor P. A. Sheppard, was in the Chair at the first symposium on ' Atmospheric aerosols.' Papers were read by Prof. L. W. Pollak on ' Photo-electric nucleus counter and its applications,' Prof. P. J. Nolan on ' The calibration of the photo-electric nucleus counter,' Mr. G. J. Day on ' The distribution of Aitken nuclei in the free atmosphere,' Dr. K. H. Stewart on ' Some optical measurements in fogs and their interpretation ' and Mr. W. G. Durbin on ' Daily aircraft sampling of condensation, freezing and chloride nuclei in January 1958.'

Prof. L. W. Pollak took the chair at the second symposium, which began with a discussion on ' Aeronautical meteorology,' opened by Mr. J. W. O'Byrne on ' Meteorological service for transatlantic flights ' and followed by Mr. P. K. Rohan on ' Some aspects of aviation route weather over the North Atlantic.'

The second half of the morning's discussion was opened by Mr. E. Knighting on ' The movement of depressions,' followed by Mr. R. P. W. Lewis on ' Jet streams.'

An exhibition of work by early Irish meteorologists was on view in the Library of Trinity College during the meeting.

Social events included a visit on Thursday afternoon to the Guinness Brewery preceded by luncheon, by kind invitation of the Managing Director, Mr. C. K. Mill, to whom thanks were expressed by the President.

On that evening some of the party made a tour of Prof. Pollak's Department in the School of Cosmic Physics, while others at the invitation of Dr. M. Doporto, Director of the Irish Meteorological Service, visited the Meteorological Office at Dublin Airport.

On Friday afternoon there was a very enjoyable excursion by coach to Glendalough and Avoca, when Mr. E. Knighting thanked the hosts for the excellent local arrangements. On Friday evening a reception was held by the President of University College, Dublin.

The Council record their sincere gratitude to Mr. R. F. Jones, who acted as co-ordinator, and to Prof. L. W. Pollak and his colleagues on the local Committee who contributed so much to the success of the meeting.

A fuller account of the Meeting and the papers read has been published in the September issue of *Weather*.

Summer Visit, 9 July 1958

The Summer Visit of the Society was held on 9 July 1958 at the Rothamsted Experimental Station, Harpenden, Herts, by kind invitation of the Director, Sir William Ogg.

Some 70 Members and guests attended and were given an opportunity of inspecting the meteorological aspects of the work in the Physics Department, in two Biological departments, and in the M.A.F.F. Plant Pathology Division housed on the Station. A short tour of the farm included a visit to the historic Broadbalk wheat plots.

At the close of the Meeting Cdr. F. H. W. Green expressed appreciation of a most enjoyable visit and of the excellent arrangements made by Dr. H. L. Penman, Secretary, in his capacity as Head of the Physics Department of the Station and by his many colleagues.

Reviews

Physikalische Dynamik der Atmosphäre. By A. Defant and F. Defant. Frankfurt am Main (Akademische Verlag., M.B.H.), 1958. Pp. xv, 527; 139 Figs., 28 Tables, 60 DM.

Statics of the atmosphere; general thermodynamics of dry and moist air; variations of state in vertical air columns, stability and instability problems; radiation; kinematics; the equations of motion on the rotating earth; simple stationary atmospheric horizontal flow; circulation and vorticity; discontinuity surfaces; turbulence and exchange (Austausch); friction and turbulence in the atmosphere; atmospheric energetics; the general circulation of the atmosphere; numerical weather forecasting; atmospheric tides and tide-like phenomena; the fifteen title headings indicate the scope of this new text-book on atmospheric dynamics.

The approach is fundamental, so providing a work suited to a first university course. The standard set by the mathematical analysis, is that now almost universally adopted in general meteorological text-books; free use of the partial differential equations of fluid dynamics with vector notation where helpful but with only limited and rather elementary manipulations. Meteorology has not yet reached the stage where large numbers of specific problems can be solved by more sophisticated methods of mathematical analysis, although there is a clear tendency in this direction, and we do not yet see the lists of examination questions which grace the chapters of many university texts in mathematical physics.

A striking feature of the work is the great preponderance of pre-war references throughout the first twelve chapters; in several there is no reference which could now be called modern. This to a large extent reflects the fact that the fundamentals of the science are satisfactorily and permanently established, but in at least some of these early chapters it is wrong to give the impression that little from post-war work needs attention even in an introductory text. Radiation, for example, has been extensively studied; turbulence has been a very live subject; energy in the atmosphere has provoked many papers which surely should not be ignored. Another comment which I think is fair concerns the chapter on surfaces of discontinuity, where the latest reference is to Bergeron 1934. In so far as surfaces of discontinuity have relevance to meteorology it is mainly in connection with fronts and one looked for some comment on real fronts which bear little resemblance to the ideal Margules surface.

When, however, we come to the very long chapter on the general circulation, some 180 pages, including much descriptive matter on synoptic disturbances as well as on climatological means, we have the feeling of entering modern meteorology, with Palmén, Rossby and Riehl justifiably accorded much attention; and the chapter on numerical weather forecasting is of necessity modern although merely introductory.

Although, throughout, the work is eclectic and we may look in vain for recognition of authors to whom other text-book writers have given prominence, the point is not all-important. The first requirement of a general introductory text-book is to provide a useful course of study and not to give recognition in swollen bibliographies to every significant contributor. I think the authors have compiled a book which meets this requirement. It contains a few things which cannot readily be found in available English texts and, if it is not outstanding in any way, it is well written and excellently printed and will be a useful addition to the library of anyone who has a liking for German and can afford to collect more than bare necessities.

R.C.S.

Atlas of five-day normal sea-level pressure charts for the northern hemisphere. By J. F. Lahey, R. A. Bryson and E. W. Wahl. University of Wisconsin Press, 1957. One page of text, 73 pp. (24 × 9½ in.) containing 219 maps. $4.

For too long, pressure changes accompanying the normal seasonal development of the general circulation have been considered almost solely in terms of monthly normals, though the conventional division of the year into months may not correspond to all important phases of the normal seasonal progression and thirty days may be too gross a unit. The appearance of an atlas of 5-day ' normal ' pressures for successive non-overlapping 5-day periods throughout the year is to be welcomed as serving an outstanding need.

M.S.L. isobars are shown at 3 mb intervals covering the zone 30° to 65°N. Additionally, standard deviation of the 5-day mean is mapped for each period by means of 1 mb isobars. Another set of maps shows the ' normal ' change between the averages for one period and the next. The use of this last set of maps is far from obvious, since the greatest ' normal ' changes indicated are no more than three-quarters of the standard deviations and mostly appreciably less than this.

Some explanatory text covering the main outlines of interpretation would have been helpful as a guard against facile use of the maps for making forecasts. How normal is the 5-day ' normal ' ? The standard deviations, running up to 17 or 18 mb at a few points in winter (compared with a maximum of 10 mb for monthly normals), help in this assessment.

The atlas was actually prepared as an aid to research with the study of singularities specially in view, and it is as a climatological research tool that such maps are properly used.

Several of the choices made in producing the atlas have left their mark upon the work. It is a great virtue that all parts of the map relate to the same years; the choice of a broken period 1907-13, 1925-37 was inevitable if Siberia was to be covered, but will make the maps less interesting in connexion with the search for climatic changes when other epochs can be compared. (For Europe and the North Atlantic comparable maps for 1890-1909 were drawn by von Elsner many years ago). The present atlas has adhered rigidly to the 65°N limit, although data were available north of this in the European sector, and in at least one case this has led to a misleading isobar pattern rather near the British Isles. On the other hand, the

whole Pacific Ocean sector is covered in spite of relative weakness of the basic data coverage such as should surely be indicated by broken isopleths or by some word in the introduction.

This atlas will be a valued possession, subject to the reservations mentioned, in research units where the large-scale circulation is studied, but assuredly most users will wish the maps had been printed in a style which at least included latitude and longitude lines over the oceans.

H.H.L.

Farming weather. By L. P. Smith. Edinburgh (T. Nelson and Sons), 1958. Pp. ix, 208; 38 Figs.; 8 Plates; Tables. 15s.

The author is well qualified to write this work which is so much needed: he has been Head of the Agricultural Branch of the Meteorological Office since 1949 and was holder of a Nuffield Travelling Fellowship in Applied Meteorology in 1954-55.

It is acknowledged to be difficult to write a book with a scientific content which will appeal to the general reader and also embrace the subject sufficiently to make it useful to the expert practitioner. L. P. Smith has succeeded well in this task, thanks to a lively style and the use of helpful analogies where the scientific concept is at all difficult. The work is not intended for the research worker but contains as much agricultural meteorology as the experimentalist and members of an agricultural advisory service are likely to need within the limits of present-day knowledge.

The book is divided into three Parts. Part I – ' The weather,' consisting of twelve chapters, provides an introduction to the subject and applies meteorology to farming tactics and day-to-day operations. The book does not instruct the farmer what to do but helps him to think more clearly. It explains to him much that might otherwise be mysterious in soil temperature, soil moisture content, plant reactions to light and temperature, radiation frosts and wind effects on vegetation. He will learn how the meteorologist has come to his aid through co-operation with the plant pathologist, the plant physiologist, the entomologist and the soil physicist. The farmer must decide for himself, however, whether he needs irrigation plant to compensate for rainfall deficiency or whether to instal heaters or some other protection against frost in his orchards, or tripods for hay drying and so on. It seems to the reviewer that almost too great pains have been taken to avoid giving advice on farming equipment and methods.

Part II – ' The climate ' – comprises seven chapters with titles : ' The approach to climate,' ' temperature climate,' ' humidity climate,' ' rainfall climate,' ' wind and tempest, frost and snow ' and finally ' the changing climate.' The climate discussed is almost exclusively that of the British Isles. 13 outline maps with isopleths are included. Instead of the customary isotherms and isohyetals we have isochrones illustrating the beginning and end of the ' growing season ' (threshold value 42°F), isopleths of average summer mid-day humidity, of average potential transpiration and of average summer soil moisture deficit. We are indebted to the author for some of the original work in this Part.

It may surprise many that Part III – ' The forecaster ' is a mere five chapters. Does this indicate that knowing how to take advantage of the weather is as important to the farmer as knowing what weather to expect ? The idea is still prevalent among farmers that the only assistance he wants from the meteorologist is his forecasts; this idea may die hard but L. P. Smith's book should go some way towards killing it.

The meteorologist, unfamiliar with agriculture, may be surprised to learn from this work how extensive has become the application of his science to farming.

M.J.S.

Linkes Meteorologisches Taschenbuch. Neue Ausgabe Bd. III. Ed. F. Baur. Leipzig (Geest and Portig), 1957. Pp. xvi, 441. 117 Figs.; 172 Tables. 28 DM.

This is the third and last volume of the new edition of ' Linke,' subtitled ' Hilfsmittel des beobachtenden Meteorologen.' It takes a curious form, comprising six essays on oddly chosen subjects, some of them not too obviously connected with observational meteorology as it is understood in this country.

F. Möller and F. Volz open with an excellent summary of the ' bases and formulae of radiation research ' (40 pages). R. Meyer follows with 20 pages of descriptive matter on some aspects of atmospheric optics. H. Israel and H. Dolezalek then review the subject of measurements in atmospheric electricity. They occupy 120 pages, and give an exhaustive bibliography, but much of their matter is of historic interest only. E. Flach then has 90 pages on the foundations of bioclimatology. This is a subject to a large extent ignored by meteorologists outside the continent of Europe. Dr. Flach sees the need in the first place to defend his subject as a scientific discipline, and proceeds to what seems to be a reasonably objective statement of its aims and methods. This is a valuable appreciation for English readers who like myself may have found the original papers heavy going. I was not convinced by it that I had missed anything of scientific value; others may form a different opinion.

The remaining two articles, on surface measurements and instruments, and on aerological measurements, by W. Grandmann and F. Woelfle respectively, are more conventional. The latter article contains a very competent summary of the principles of radiosonde design, and a useful if uncritical description of sondes currently in use.

The volume as a whole is a rather disappointing sequel to the very fine second part. There is little reason why it should be bought by anyone whose native language is not German.

G.D.R.

The ionosphere. By Karl Rawer (translated from the German by Ludwig Katz). London (Crosby, Lockwood), 1958. Pp. 202; 72 Figs.; 7 Tables. £2 2s. 0d.

The original German edition of this book which appeared in 1952 was reviewed by J. A. Ratcliffe (*Q.J.*, **79**, 1953, p. 309). The text has not been changed, so that little need be said about the material except to agree with Mr. Ratcliffe that the book meets a very great need though the material is partly controversial and the construction is not above criticism.

The translation is adequate for understanding, but that is all. English is clearly not the mother tongue of the translator, and apart from oddities of syntax and usage, one finds non-trivial mis-translations, e.g., 'shell method' for 'grenade method' and occasional German phrases, e.g., 'Vgl. etwa' and 'siehe' in the Index.

R.M.G.

Natural aerodynamics. By R. S. Scorer. London (Pergamon Press), 1958. Pp. xi, 312; Figs; Tables; 22 Plates. 60s.

In this book some of the basic concepts of fluid dynamics are discussed in relation to problems associated with air pollution, mountain waves, and thermals. A brief derivation is given of Bernoulli's Theorem and the equations of fluid motion, and the concepts of vorticity, viscosity, boundary layer, turbulence and buoyancy are described. An attempt is made to minimize the use of mathematics and to emphasize the physics, the idea presumably being that the work may then be intelligible to non-mathematicians.

There is no doubt that observation of the flow of smoke and water in natural surroundings together with the kind of stimulating discussion given in this book would indeed form a working basis for those interested in the study of natural aerodynamics as a hobby. In this context the sections concerned with the soaring of birds, the motion of bubbles of air in water, the capture of small bubbles in the wake of larger ones, the behaviour of buoyant plumes, and the production of mountain waves are particularly interesting.

The problem of turbulence is of inherent interest in natural flow and the rather brief treatment given is a useful one, the difficult problem of defining mean values of wind velocity, concentration of smoke, etc., being rightly emphasized. One basic experimental result is, however, omitted : this is that if we measure smoke concentration continuously with time in neutral stability conditions, then an approximate, stationary, average distribution is established after about three minutes and Sutton's formulae are then applicable. For longer sampling periods the time change in the averaged values is very slow.

Finally, despite the author's suggestion, it should be noted that anyone reading this book seriously should first be reasonably familiar with the elements of vector analysis, calculus and fluid dynamics. It will be of assistance mainly to those who, having already studied fluid dynamics in a mathematics course, wish to work on problems of air pollution or mountain waves. The author has undoubtedly injected into the book some of his enthusiasm for his subject, but it is regrettable that it is somewhat marred towards the end by references to metaphysics, which the reviewer feels are somewhat misplaced in a work of this nature, absence of proper punctuation in places, and some ambiguous sentences. The standard of production is very high, but the price is also high for a book of this kind.

D.R.D.

An introduction to Fourier analysis and generalized functions. By M. J. Lighthill. Cambridge (University Press), 1958. Pp. viii, 79. 17s. 6d.

This is quite a delightful little book on Fourier integrals and series viewed in a different way from that which we usually encounter in the standard texts. The difficulties of determining under what conditions a function has a Fourier integral or series are largely removed by considering the conditions for a class of functions, here called generalized functions; a study of some particular generalized functions leads to a calculus not unlike that of residues, which enables Fourier transforms of rational functions to be written down directly. Asymptotic values of integrals of the Fourier type are treated, and finally Fourier series are linked with Fourier integrals of generalized functions.

The ideas are not difficult to grasp but the reader may have some difficulty with the new vocabulary – 'good function,' 'the smear function,' 'the unitary function.' It is stimulating and written in a style which is easy to read and understand. Meteorologists who have to use Fourier methods would do well to buy and read this book. It is quite cheap by present standards.

E.K.

Die Höhenwinde über dem Tropischen und Südlichen Atlantischen Ozean; A. Allgemeine statistische Bearbeitung der Höhenwindmessungen. By Erich Kuhlbrodt ('Meteor' Expedition, 1925-1927, Bd. XVI, Teil 1) Berlin (W. de Gruyter), 1958. Pp. 91 (13 pages text, 14 Figs.; 26 Tables). 37.50 DM.

The pilot-balloon ascents made from the Research Ship *Meteor* during the expedition of 1925-27 were fully described by Kuhlbrodt and Reger in Vol. XV of the *Meteor Reports*, published in 1933.

Professor Kuhlbrodt has now made a statistical analysis of the *Meteor* upper-wind observations together with those made by the Deutsche Seewarte up to 1939. The results are fully presented in Figures and Tables: they provide a welcome addition to our knowledge of the windfield over the Tropical and South Atlantic.

H.C.

Three steps to victory. By Sir Robert Watson-Watt. London (Odhams Press), 1958. Pp. 480; Figs.; Plates. 30s.

Although the author says he was never a diaryman and writes with next to no personal record, this autobiography in 76 chapters is a book to be read and not summarized, so full is it of facts and opinions presented with the author's characteristic thoroughness. Reading through the three pages of chapter headings one finds Watson-Watt in lighter mood, but in the book itself he is in deadly earnest. At first the style is strange, but the story emerges through the noise, impelled by Watson-Watt's driving power, from the early days of thunderstorm recording at Farnborough in World War I via the request for a death ray to the trials and triumphs of radiolocation in World War II.

For us the story really starts in 1915 at *the* Branch Meteorological Office at South Farnborough (there was then only one B.M.O.) with J. S. Dines, Cave and Watson-Watt and, on the radio side of the ' Factory,' Richard Whiddington, later Professor of Physics at the University of Leeds and President of the Physical Society. In connection with the daily sending up of a pilot balloon, 25 years before the equally frequent use of radiosondes, these meteorologists occupied their spare time in measuring the relation between the rate of ascent and the dimensions of a balloon in a series of experiments under the echoing dome of an empty Royal Albert Hall.

To one who joined the meteorological service about the same time as the author and left it about the time of publication of this book over forty years later, there is in it a long succession of friendly names in science and in the Royal Air Force. But one feels, with the author, that one friendship is somewhat strained when one is poignantly reminded of a period in the twenties when it was fortunately (!) unnecessary to spend much more than one-tenth of the £2,000 allowed for a whole year's research in the British Meteorological Service; we know just how the author felt. It was about this time that C. E. P. Brooks worked out (was it on the back of an old envelope ?) that on an average over the whole world there are about 100 lightning flashes every second and thus comforted the thunderstorm workers who were finding a surprisingly large number of *X*'s.

The major part of the book is of special interest to radio enthusiasts and generally it shows researchers how to reach their goal; it is good value at the price. Clearly in accord with the Duke of Wellington's ideas, it is a sermon on the text, ' By God ! I don't think it would have been done if I had not been there.'

R.M.P.

Your T.V. weather map. By J. J. Higgins. Glasgow (Blackie and Son), 1958. Pp. 90; 38 Figs. 9s. 6d.

This is an elementary little book which was written primarily to enable the reader to make greater use of official weather forecasts. The aim is to help the layman to associate the sequence of weather he experiences with the weather maps shown on his television screen or in the newspapers.

The difficulty in writing this kind of book is to decide how simple to make it. Too advanced a treatment defeats the object of the book. On the other hand, an over-simplified and sketchy description irritates the reader who expected something to get his teeth into and finds the meal underdone. For the reader with nine shillings and sixpence to spend on his introduction to forecasting the author ought perhaps to have aimed a little higher. Nevertheless, the writing is lively, the chapters are short and compact and the reproduction is elegant, so the reader's interest is maintained to the end.

The author gives a fairly detailed picture of the structure of idealized fronts and the sequence of weather in a depression. The greater part of the book is written around movements of synoptic features on weather maps, though there are excursions into such subjects as thunderstorms and ' smog.' This emphasis inevitably leaves the impression that forecasting the correct positions of fronts contributes more to an accurate forecast than is in fact the case. This is perhaps a limitation of any book that sets out to be largely descriptive and attempts no explanation, however inadequate, of most atmospheric processes.

It is a little disappointing that forecasting is presented as a somewhat static science of no great complexity. There is little in this book that could not have been written 25 years ago, and the writer might well have added to its interest by including something about upper air observations, the uses of radar, how thunderstorms are located, the discovery of the jet stream and the coming of the machine age in forecasting. A popularized account of any of these could be quite absorbing, but none is even mentioned.

The opinion of two acquaintances who are not meteorologists was that *Your T.V. weather map* was very readable and helpful. Judging by what the book includes rather than what it omits this assessment seems a very fair one.

C.J.B.

Meteorological thermometers. British Standard 692. London (British Standards Institution), 2nd Revision, 1958. Pp. 18; 5 Figs.; 3 Tables. 4s. 6d.

The British Standards Institution is financed by subscriptions from firms, trade associations, professional institutions, etc., and by a Government grant. The principal objects of the Institution are to co-ordinate the efforts of producers and users for the improvement, standardization and simplification of engineering and industrial materials; to simplify production and distribution; to eliminate the waste of time and material

involved in the production of an unnecessary variety of patterns and sizes of articles for one and the same purpose; to set up standards of quality and dimensions, and to promote the general adoption of British Standards. It is the policy of Government Departments to adopt British Standard specifications whenever this is practicable.

A British Standard for Meteorological Thermometers was first published in 1936, based on Meteorological Office specifications, and was revised in 1951. The most important change in this 2nd revision is the inclusion of a parallel series graduated in degrees Celsius. Less important changes are the exclusion again of unsheathed thermometers (first included in a war emergency addendum); and a small change in the safety chamber requirements. (The reason given for the exclusion of unsheathed thermometers is that they are very little used).

A point of interest is the use throughout of the abbreviation ' deg F (or C) ' instead of ' °F (or C) ' whenever an *interval* of temperature is denoted – though the effect is spoilt by its incorrect use, in clause 12, to denote an actual temperature.

R.F.

The climate of New Zealand. By B. J. Garnier. London (Edward Arnold), 1958. Pp. 191; 72 Figs.; 50 Tables; 12 Plates. 70s.

This attractively produced volume contains a most lively and authoritative description of the climate of New Zealand, supplemented by enough tables and diagrams to make the work of practical use without reducing it to a dry catalogue of facts. Throughout the book the various weather types (which constitute the climate) are described in relation to the general atmospheric circulation, the synoptic pattern and topography. Many of the diagrams are synoptic charts typical of the types of weather and the work could be quite properly regarded as descriptive synoptic climatology. The twelve beautifully produced plates consisting mostly of panoramic views are a most attractive feature of the book and add life and interest to the well-written text.

The first four chapters are a general account of the climate, geography and physical character. The climate combines rather more sunshine than occurs in southern England with ample rainfall and general windiness. The frequent and more or less regular passage of cyclonic troughs and intervening ridges with alternations of tropical and polar air are associated with markedly changeable weather. The fact that the air masses are maritime in origin precludes excessively warm and wintry spells such as are experienced in England.

The remainder of the book deals in separate chapters with more detailed accounts of the climate of the nine regions into which the author divides the country. These are supplemented by some general climatic tables at the end of the book, but for full climatic statistics it would be necessary to consult the official summaries. The climate is described in relation to vegetation and the main industry of the island, agriculture; its influence, actual or potential on manufacturing industries, building practices and so on, are hardly touched upon. A small defect is the absence of a key map showing the numerous places and geographical features mentioned in the text.

Altogether a most pleasing volume, it would find a place on the home bookshelf equally well as in the technical library.

A.G.F.

Geography of North America. By G. J. Miller, A. E. Parkins and B. Hudgins. London (Chapman and Hall), 3rd Edition, 1954. Pp. xi, 664; 290 Figs.; 35 Tables. 60s.

The story of European man's occupancy of the North American continent is largely a record of his exploitation of superb natural resources, and the people of the U.S.A. and Canada are still profiting greatly from material assets that assure opportunities for maintaining high standards. It is the purpose of this book to trace, in interpreting the present-day geography of North America, man's use of the natural resources available to him. Among the several that can be identified, the climatic endowment is the one of most interest to readers of the *Quarterly Journal* and the answer to the question ' what use has man made of the climatic asset and in which respects has it presented him with difficulties ? ' is not to be found in any neatly organized analysis. Instead, it is to be found throughout the book and with different regional emphases. Only in the section on the Western States, where human adjustment to climatic conditions is clear and climatic influences approach controls, is a chapter devoted to climate. In other sectional treatments the climate is incidental to the other phases of the geography. The result is that climate is treated in simple descriptive terms and with emphasis on those aspects that have broad agricultural significance, viz., length of growing season, amount and seasonal distribution of rainfall, etc. It would be easy to criticize this use of a superficial descriptive climatology with no hint of a genetic appreciation through dynamic climatology, but the authors claim only that the book is a basic text, an introductory course contributing to a general education and providing a starting point for the specialist geographer's study of the continent. To the general reader it does offer the salutary reminder that what man does is not to be understood independently of the qualities of the earth on which he lives.

N.P.

The climate of central Canada. By W. G. Kendrew and B. W. Currie. Ottawa (Queen's Printer and Controller of Stationery), 1955. Pp. ix, 194; 52 Figs.; 212 Tables. $1.

The climate of British Columbia and the Yukon Territory. By W. G. Kendrew and D. Kerr. Ottawa (Queen's Printer and Controller of Stationery), 1955. Pp. x, 222; 54 Figs.; 103 Tables; 8 photographs. $1.

To meet the Department of Transport's need for a comprehensive study of the climate of Canada, initial detailed reports of climatic conditions in the several parts of the country were produced by specialist investigators. For publication for general use, the detailed reports needed to be condensed and combined and these two volumes, the product of such rewriting by W. G. Kendrew, give evidence of the soundness and usefulness of the series planned to cover the greater part of Canada. The arrangement of treatment in the two volumes is basically similar in that general considerations of relief, major climatic regions and a survey of the coverage of meteorological information preface the systematic climatic account. In this the topics are the standard ones that would be expected, namely, pressure systems, winds, air masses and frontal zones; temperature and humidity; clouds and sunshine; precipitation; visibility and, in central Canada, local climate. All sections are well illustrated by clearly drawn maps and diagrams and are fully supported by tables within the text and by supplementary climatological tables. To list the sections under which the material is organized in order to indicate the coverage of the reports gives the author less than his due, for it is only by reading the text that one appreciates that the easily readable narrative enhances considerably the value of the readily available data. These condensations of a vast amount of data will be essential reference volumes for anyone concerned with Canadian climates but the skill of the author in providing a text that is an attractive and illuminating commentary on these meaningful climatic data will make these volumes also invitingly readable by the general public for whom they were intended.

N.P.

Obituary Notices

Dr. C. F. BROOKS

Charles Franklin Brooks, Emeritus Professor of Meteorology at Harvard University, died suddenly on 8 January 1958, in his 67th year. Although he had retired last year from the directorship of Harvard's Blue Hill Meteorological Observatory, he had remained active in meteorological affairs; on the morning of his death he had made numerous samplings and measurements of the fresh snowfall. Dr. Brooks was born in St. Paul, Minnesota. The severe summer thunderstorms which occur in that part of the country impressed him during his youth and served to spark his interest in the weather. His university education was pursued at Harvard, where he received the A.B. in 1911, the M.A. in 1912, and the Ph.D. in 1914. His Ph.D. in meteorology was the second to be awarded in the U.S.A. He went to Washington, D.C., to work on the *Atlas of American Agriculture*. Then he became instructor in geography at Yale University (1915-18). Later he served as instructor in meteorology for the U.S. Army Signal Service at College Station, Texas. After this he returned to Washington to edit the Weather Bureau's *Monthly Weather Review* (1918-21). For the next decade he taught meteorology and climatology in the Graduate School of Geography at Clark University. The American Meteorological Society was founded in 1919 largely through the personal energy and enthusiasm of Dr. Brooks. He was elected secretary at the first meeting and served in that office until 1954. He edited the society's *Bulletin* during the first 19 years. In 1931 he was appointed Professor of Meteorology at Harvard and Director of the Blue Hill Meteorological Observatory. In 1940 the Weather Bureau Solar Radiation Field Testing Unit was moved from Washington, D.C., to Blue Hill at Dr. Brooks's suggestion, because the new location provided better exposure and proximity to other laboratories experimenting with the uses of solar energy. In 1932 Dr. Brooks was instrumental in re-establishing a weather observatory at the top of Mount Washington. He became its president and engaged actively in this work until his death. From the first few years of the Mount Washington observations he was able to illustrate their value to the Weather Bureau as aids in forecasting storms and cold waves. This, subsequently, led to the Bureau's support of the observational programme. Liquid-water content and icing rate of clouds were first routinely observed on the mountain. This work, always heartily endorsed by Dr. Brooks, led to many experiments in cloud physics and to the collecting of valuable data for the aircraft and tower industry, the latter having developed with the need for tall television and microwave towers.

Dr. Brooks's chief scientific studies were on weather and climate. At Blue Hill he prepared, jointly with A. J. Connor, basic climatic maps of North America which are still considered the best available for the area. In the 1930's he set up a network of special weather stations over New England for the study of eclipse meteorology. Under his direction the American chronometric radiosonde was developed at Blue Hill, and, for a time, used routinely by the U.S. Weather Bureau and other countries. He studied the deficiencies of precipitation gauges, and designed and put into use new windshields which provide better ' catch ' during windy storms of rain or snow. His interpretation of clouds in terms of stability and general pattern of the atmosphere is well known among meteorologists. His knowledge of clouds, instruments, and methods of observation led the International Meteorological Organization, which later became the World Meteorological Organization, to utilize his experience in these matters in the late 1940's and early 1950's. He served this organization as a member of the committee which prepared the *International Cloud Atlas*. In 1953 he participated in the meetings of the Commission on Instruments and Methods of Observation of the World Meteorological Organization.

Dr. Brooks was a Life Fellow of the Royal Meteorological Society (elected in 1920), a member of many scientific organizations in the United States, and, at one time, President of the Association of American Geographers. He served with the U.S. Quartermaster Corps, the Office of Scientific Research and Development, and the National Research Council. His book, *Why the Weather*, published in 1921 and revised and enlarged in 1935, has helped to inspire many to become meteorologists. After the founding of the Lowell Broadcasting Council and the establishment of its radio transmitter on Blue Hill, he gave weekly talks on current events, both local and international, in the ever-expanding field of meteorology. Dr. Brooks showed a warm sympathy for all men, and gave of his time and energy unstintingly, day or night, to all who sought his advice – regardless of their academic degrees or lack of them.

J.H.C.

Captain C. ADAM

Captain Charles Adam, of Ellesmere, Salop, who was elected into this Society in 1933 and had held Life Fellowship since 1940, died on 18 January 1958. Educated at Rossall, he interrupted a business career soon after the outbreak of the 1914-18 war to serve abroad with a commission in the King's Liverpool Regiment. He was severely wounded at the Battle of the Somme and was later invalided out of the Army. Subsequently he concerned himself in particular with the British Legion and with the Scout Movement, for which he was ultimately appointed deputy commissioner for Shropshire. He also took a prominent part in public life, becoming a J.P. in 1944 and deputy lieutenant for Shropshire in 1956. As an officer in the Observer Corps during the 1939-45 war he was allocated responsibility for an area extending from his home county into Staffordshire. Captain Adam's scientific interests, developed in his schooldays, led him to observational meteorology while still a young man. For nearly 30 years he had taken regular daily measurements of rainfall at Ellesmere and contributed summaries of his records for publication by the Meteorological Office.

E.L.H.

Miss VERA M. HUCKLE, B.Sc.

Miss Vera May Huckle died on 19 May 1958 aged 27 years. During the previous 18 months she had been ill for much of the time and had spent long periods in hospital at Hammersmith and Bedford.

Miss Huckle entered the Meteorological Office in October 1952, after gaining a B.Sc. (Special) in Mathematics at Bedford College, London. In her work at the Napier Shaw Laboratory, Dunstable, she was a member of the small team which during the early 1950's embarked vigorously on research into numerical methods of treating the dynamical equations of meteorology. She played a significant part in this work, particularly in the programming of lengthy calculations for solution by electronic computer. She was co-author with F. H. Bushby of two papers related to numerical forecasting published in the *Quarterly Journal*. It is saddening that she should not have lived to see the milestone which will shortly be reached in this work with the installation of the Meteorological Office electronic computer at Dunstable. She will be missed by her colleagues, not only for the loss of her undoubted talents, but also because her quiet sincerity and devotion to duty made her well liked by all who came in contact with her. Miss Huckle was elected a Fellow of the Society on 20 November 1957.

F.H.B.

Sir ERNEST MACLAGAN WEDDERBURN, O.B.E., D.L., M.A., LL.D., D.Sc., F.R.S.E.

By the death of Sir Ernest Wedderburn the last link with the Council of the Scottish Meteorological Society is broken. Ernest Maclagan Wedderburn joined the Scottish Society in 1908 and almost immediately was elected joint honorary secretary with R. T. Omond. He had been attracted to geophysical studies during his university course in arts and law at Edinburgh by the enthusiasm of the Professor of Mathematics, George Chrystal, who was at this time engaged in organizing a programme of observations of seiches in lakes in connection with Sir John Murray's survey of Scottish lochs. While still an undergraduate and later while establishing himself in his chosen profession of law, he spent every summer holiday between 1901 and 1913 making observations on Loch Ness, Loch Garry and Loch Earn and published a series of papers in the Transactions of the Royal Society of Edinburgh and the Scottish Meteorological Society. Even before he became a Writer to the Signet (W.S.), he was already a Fellow of the Royal Society of Edinburgh and later he was awarded the Makdougall Brisbane Prize of that Society. The scientific work was to him a hobby, pursued in holidays and spare time; he was meanwhile acquiring a considerable reputation in the legal profession and had begun to concentrate his interest in conveyancing.

By 1914 he had become a partner in one of the largest Scottish legal firms, but soon after the outbreak of war he volunteered for service and became Captain-in-charge of the meteorological unit, Royal Engineers, attached to the forces in the Eastern Mediterranean. For his distinguished services at Gallipoli and in Macedonia and for his work at Salonika on the application of meteorology to gunnery, further developed when he became Assistant Superintendent of Experiments at Shoeburyness, he was awarded the O.B.E. (Military Division). On demobilization in 1919, he was elected to the chair of conveyancing at Edinburgh where he introduced ' the simple but then quite new practice of giving his classes instruction and practice in the preparation of the forms of deeds with which they would later be confronted. This was at the time something of a revolution in methods of teaching.' In 1935, he resigned his chair on being appointed Deputy Keeper of the Signet (D.K.S.), the head of the oldest body of solicitors in Scotland, and received his knighthood. During his tenure of the office of D.K.S. from 1935 to 1953, he played the leading part in the government of the legal profession in Scotland. His record of public service, much of it performed so unobtrusively that it is only now becoming generally known, is equally impressive. To mention but one of his activities, he was deeply interested in Boy Scouts, serving as a scoutmaster in the early years of scouting and later as county secretary and county commissioner, and received the highest award for services to the movement, the Silver Wolf.

The establishment of the Scottish Centre of our Society in 1946 gave Sir Ernest great pleasure and he seldom missed a meeting. It had fallen to him as honorary secretary to make the arrangements in 1920-21 for the amalgamation of the Scottish Meteorological Society with the Royal Meteorological Society. He later confessed to a feeling ' that I was aiding and abetting a murder when the old Society lost its identity on being incorporated in the Royal Meteorological Society. That it should be resuscitated lifts a load off my conscience ' (*Q.J.*, **74**, p. 233). He served as Scottish Chairman and Vice-president from 1949 to 1952 and again during 1955, the year of the centenary of the foundation of the Scottish Society. He was an admirable chairman and councillor, conducting the business quietly and efficiently. Kindly, unassuming and generous, he was a man of great intellect and shrewdness. Sir Ernest rendered signal service to the Scottish Meteorological Society and later to the Royal Meteorological Society and the memory of that service will not readily be forgotten. His death at the age of 74 on 3 June 1958 after a short illness came as a shock to us all, for he carried his years lightly. He is survived by Lady Wedderburn, two sons and a daughter.

The Society was represented at the memorial service held on 12 June, by Mr. D. H. McIntosh, Secretary of the Scottish Centre.

J.P.

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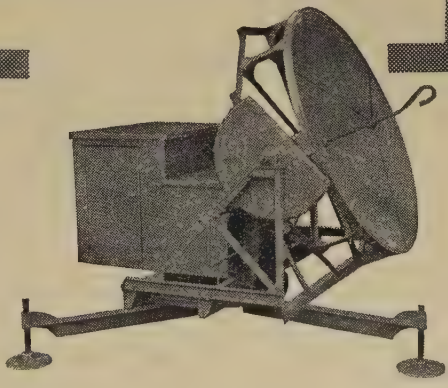

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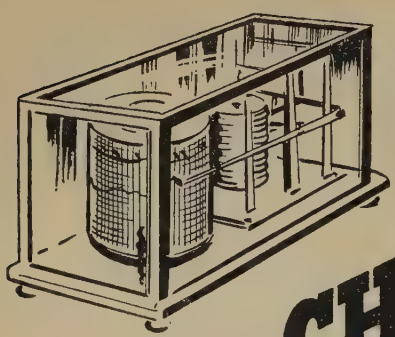

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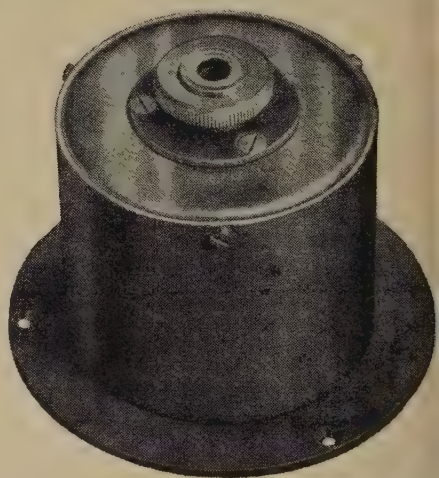

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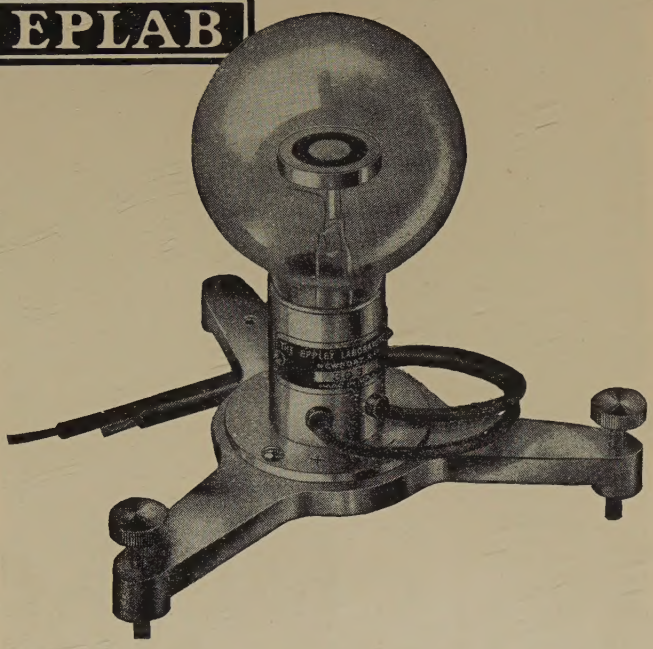

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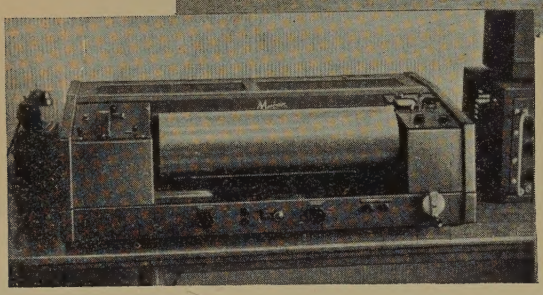

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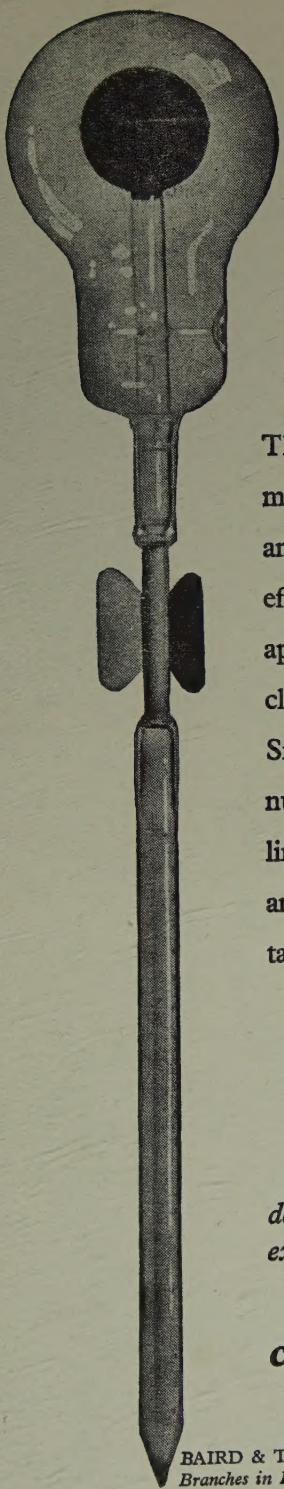

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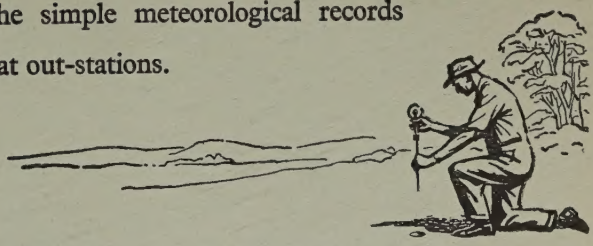

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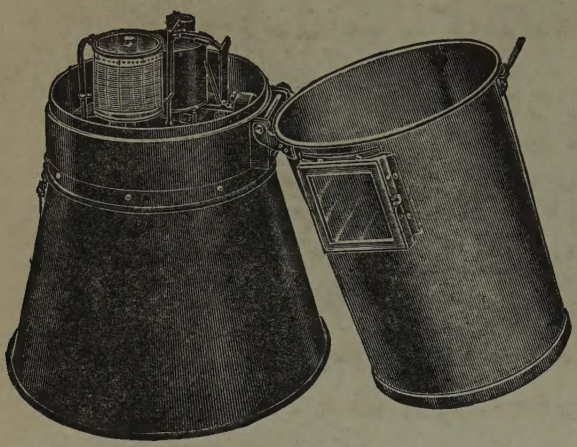

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